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Water Resources
Report 10

Hydrogeology and Ground Water Model of the Blue Springs Creek IHD Representative Drainage Basin

INTERNATIONAL HYDROLOGICAL DECADE
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Ontario

WATER RESOURCES
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Hydrogeology and Ground Water Model of the Blue Springs Creek IHD Representative Drainage Basin

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MINISTRY OF THE ENVIRONMENT
Water Resources Branch

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PREFACE

At its thirteenth session, held in Paris in October-November, 1964, the General Conference of the United Nations Educational, Scientific and Cultural Organization (UNESCO) set up the Co-ordinating Council of the International Hydrological Decade. The work of the Council lead to an international program in the field of hydrology. The overall objectives were to accelerate the study of water resources and the regimen of waters with a view to their rational management in the interest of mankind, to make known the needs for hydrological research and education in all countries, and to improve each country's ability to evaluate its water resources and use them to the best advantage. The program thus focused on science, but gave strong consideration to utilitarian factors.

As part of the contribution to the International Hydrological Decade program, the Province of Ontario investigated the water resources and physical conditions in five drainage basins in Southern Ontario, each being representative of different conditions common in the province. The hilly moraine and mild karst topography of the Blue Springs Creek basin was selected as one of the representative areas. This report describes the hydrogeologic characteristics of the basin as deduced from data collected during the International Hydrological Decade, 1965-1974. The study was completed by the Water Resources Branch of the Ontario Ministry of the Environment.

G.H. Mills
G.H. Mills, Director
Water Resources Branch

Toronto, February 1978

CONTENTS

FIGURES

* in pocket

TABLES

PHOTOGRAPHS

MAPS
(in pocket)

Map

ABSTRACT

A study of the occurrence, quantity and quality of the ground water was undertaken in the Blue Springs Creek basin by the Ontario Ministry of the Environment, as part of the International Hydrological Decade (IHD) Representative Basin program. The Blue Springs Creek basin was selected for study to represent a physiographic region in Southern Ontario, characterized by shallow overburden and karst limestone and dolomite bedrock topography. These conditions are generally found along the top of the Niagara Escarpment.

Field work for this study included the investigation of the surficial geology, the installation of 16 observation wells to aid in the investigation of the subsurface geology and ground water movement, streamflow monitoring, and chemical sampling of surface water and ground water systems.

The major aquifer found in the basin is the Amabel Dolomite Formation; it averages 60 feet (18 metres) in thickness, reaches a maximum thickness of 100 feet (30 metres) and outcrops over approximately 10 per cent of the basin area. The overburden material, covering the remaining 90 per cent, forms a minor aquifer, composed largely of till and kame moraine deposits with a thickness of almost 150 feet (45 metres) in the northern portion of the basin.

The Amabel Formation in the basin was found to have a median transmissivity of 3,200 IGPD/ft (550 mm^2/s,) similar to that of the overburden aquifer at 3,500 IGPD/ft (600 mm^2/s). The transmissivity values found for shallow wells in the Amabel Formation were generally higher than those found in deeper wells in the same formation.

Baseflow in the basin was found to average approximately 68 per cent of the total streamflow during the water years of 1966-69, and was equivalent to an annual ground water runoff rate of 9.8 inches/yr (250 mm/yr). The average specific yield was found to vary from 7.2 to 12.7 per cent in various sub-basins of the Blue Springs Creek basin.

Two finite difference ground water models were used to simulate the steady state and dynamic flow patterns in the basin, and to improve or verify the estimates of transmissivity, coefficient of storage, recharge and baseflow obtained by conventional techniques. Good representation of the piezometric surface was obtained with the model if the regional bedrock transmissivities used in the model were 1.5 times the median transmissivities found at wells, the storage coefficients ranged from 4 to 16 per cent and the recharge rates varied from 7 to 15 inches per year (180 to 380 mm per year) in different areas of the basin.

In examining the hydrochemistry of the flow systems, the majority of the ground water samples taken from bedrock wells indicated that the water was of the calcium-magnesium and bicarbonate-chloride sulphate type, saturated with respect to calcite and undersaturated with respect to dolomite.

The calcium to magnesium ratios of the water samples ranged from 1.31 to 4.61 with a mean of 2.10 in comparison with a mean ratio of 1.09 in the bedrock material and 2.15 in the overburden material. The higher ratios in the ground water tended to be in the recharge areas, and the lower ratios in the discharge areas. Using these results the generalized ground water flow patterns were determined, and broadly confirmed the piezometric flow patterns.

INTRODUCTION

SCOPE OF THE INVESTIGATION

The study of the hydrogeology of the Blue Springs Creek basin is one of several detailed water resources investigations being completed by the Ontario Ministry of the Environment, as part of the International Hydrological Decade (IHD) Representative Basin program (Figure 1). The University of Guelph studied the surface water hydrology as an IHD co-operative project with the Ministry of the Environment.

The major areas of study described in this report pertain to investigations of the geology, hydrogeology and hydrochemistry of the Blue Springs Creek basin. Computer models were applied to simulate the hydrogeologic conditions.

The study included field investigations of the geology, hydrogeology and hydrochemistry. Water well records maintained by the Ministry of the Environment were used to characterize the aquifers, water levels and subsurface geology in the region. The streamflows in the basin were monitored by three stage recorders maintained by the Ministry of the Environment and Environment Canada, as well as five staff gauges which were read about once a month. Ground water levels were monitored by four recorders set up on observation wells, and by approximately two monthly readings taken at four other wells. Aerial photographs obtained from the Ministry of Natural Resources were used to supplement field studies.

OBJECTIVES

The basic objectives of the study are summarized below. They are based on the 'Guidelines for Research Basin Studies' (1966), as compiled by the Canadian National Committee for the International Hydrological Decade:

1. To study the surficial and subsurface geology of the basin and its surrounding area.
2. To study the geologic conditions under which various flow systems operate in the basin.
3. To determine the hydrogeologic parameters of the aquifers in the basin.
4. To identify ground water flow systems in the basin.
5. To delineate principal areas of recharge and discharge in the basin.
6. To evaluate the total ground water discharge.
7. To determine the basin yield.
8. To estimate the regional hydrogeologic parameters of the bedrock aquifer using a ground water model.
9. To determine the relationship between water quality and water movement.

DESCRIPTION OF THE BASIN

The Blue Springs Creek basin comprises an area of approximately 30 square miles (78 km^2), bounded by latitudes 43° 30' and 43° 41' N and longitudes 80° 00' and 80° 10' W. It contains part of Wellington and Halton counties.

The basin is almost entirely rural, including only a few small settlements such as the villages of Knatchbull and Crewsons Corners; the

Figure I. Location of Blue Springs Creek basin in Southern Ontario.

larger towns of Acton, Eden Mills and Rockwood lie just outside the drainage boundary (Map 1). When considering the overall hydrologic regime of the basin, it can be considered as relatively unaffected by human activity.

The topography of the Blue Springs Creek basin (as shown on Map 1) is characterized by broad rolling hills, interrupted by small and prominent ridges and deeply incised stream valleys. The maximum elevation in the basin is about 1400 feet (425m) above mean sea level, in the northern portion of the basin. The lowest elevation is about 1040 feet (315m), at the downstream end of the basin, giving a total relief of approximately 350 feet (110m).

The basin is dissected by the perennial Blue Springs Creek, which is a tributary of the Eramosa River in the Grand River watershed. Its length is about eight miles (13 km), with an average gradient of approximately 45 feet per mile (8m per km). The creek has its main source in the morainic hills near Acton and is also supplied by a number of springs in the basin (Dickinson and Whiteley, 1970). The creek derives its name from a large spring in the centre of the basin.

The basin contains several intermittent streams (Map 1), some of which do not reach the main channel but sink into local depressions. These springs and sinking streams are further discussed in Section (2.2.4).

The climatic and hydrologic data collected in the period 1967-1971 were used for much of the hydrologic analysis, as they formed the most complete set of data that was available when the analysis was carried out. These five years were slightly wetter than the historic period as recorded at nearby long term stations. Thus the mean precipitation at Guelph during 1967-1971 was 33.7 inches compared to the 30 year (1941-1970) mean of 32.7 inches. The mean streamflow on the Speed River below Guelph during 1967-1971 was 212 cfs compared to 197 cfs for the 24 years of record (1950-1973). As the mean conditions during the years 1967-1971 were only slightly wetter than the long term mean values (3% and 8% higher for the precipitation and streamflow, respectively), the hydrological conditions during this five year period were considered to be fairly typical of the long term conditions.

PREVIOUS INVESTIGATIONS

The Paleozoic rocks underlying the area were previously mapped by Williams (1919), Caley (1941), Bolton (1957), Sanford and Quilliam (1959), Sanford (1961 and 1969), Liberty (1969), Hewitt (1972), and Telford (1976). Hydrogeologic information on the Paleozoic rocks underlying the area is contained in water well records maintained by the Ministry of the Environment, and oil and gas well records maintained by the Petroleum Resources Section, Ministry of Natural Resources. Hewitt (1964 a, b) has also discussed the limestones of the area for use as building materials and as industrial minerals.

The Pleistocene features and deposits of the area have been mapped by Chapman and Putnam (1966) and by Karrow (1968).

Agricultural soils have been mapped by Hoffman and Mathews (1963) and by Gillespie et al (1971).

The study of watershed areas contributing to runoff in the Blue Springs Creek basin, by Dickinson and Whiteley (1970), also provided significant background data for the study.

ACKNOWLEDGEMENTS

The study and report preparation were carried out under the general supervision of Mr. R. C. Hore of the Hydrology and Monitoring Section and Mr. D. N. Jeffs and Mr. F. C. Fleischer of the Water Modelling Section, whose assistance and advice were invaluable. Over the years of the study, field and office work contributions were made by various staff members, including Messrs. B. Jaffray, G. Jordan, A. Bacchus and D. Donohue of the Water Resources Branch. Useful discussions were held with associate staff members G. Funk, R. Ostry and S. Singer, during preparation of this report.

Co-operation as received from residents in the Blue Springs Creek area during geological mapping, well drilling, chemical sampling and other field work, is gratefully acknowledged. Appreciation is expressed to those residents who acted as observers collecting hydrological data for this study.

GEOLOGY

The characteristic features of the bedrock and surficial geology of the Blue Springs Creek basin are discussed in this chapter. A summary of the various units and their position in the stratigraphic column is given in Table 1. The distribution of surficial geologic units is shown on Map 2, and cross-sections of the basin geology are illustrated in Figure 3 and Figure 5.

BEDROCK GEOLOGY

Regional

The consolidated bedrock formations in southwestern Ontario consist of Paleozoic sedimentary rocks of Ordovician, Silurian and Devonian Age, resting on the Precambrian basement (Figure 2 in pocket). The Paleozoic rocks dip regionally to the southwest at about 25 feet to the mile; the oldest rocks of Precambrian Age outcrop to the northeast and the younger rocks of Devonian Age outcrop in the southwestern part of Ontario. In the area around Blue Springs Creek, the outcropping rocks are dolomites of the Amabel Formation of Middle Silurian Age (Figure 2); older rocks of Lower Silurian and Ordovician Age have been found in the subsurface in a number of oil and gas wells drilled in the area (Figure 3 in pocket).

The Silurian rocks in Southern Ontario undergo a facies change north of Hamilton, which has been attributed either to a structural barrier called the 'Algonquin Arch' (Roliff, 1954), or to the presence of two major depositional areas during Silurian times in Southern Ontario (Bolton, 1957). The Silurian rocks contain many carbonate beds, which to the north, in the Bruce Peninsula area, (Figure 2) are more dolomitic, while to the south, in the Niagara Peninsula, they are more calcareous. In the Blue Springs Creek area, the Clinton Group (Table 1) has been correlated with the more calcareous, southerly facies of the Niagara Peninsula, while the overlying Albemarle Group is more closely correlated with the dolomitic northerly facies (Bolton, 1957).

In the Blue Springs Creek area, minor warping of the Ordovician and Silurian beds occurs (Liberty, 1969). Weber (1960) considers that the Eramosa River near Rockwood, (Map 1), runs along an anticlinal axis in the Amabel Formation. The southern part of Blue Springs Creek runs along a synclinal axis in the same formation (Liberty, 1969), but this fold does not extend into the northern portion of the basin, where the Amabel is relatively unfolded.

Blue Springs Creek Basin

In the Blue Springs Creek basin, the outcropping bedrock beds are members of the Middle Silurian Amabel Formation (see Table 1 and Map 2). The Upper Ordovician and Lower Silurian beds, however, also play a significant part in the hydrogeologic characteristics of the basin (Figure 5 in pocket). In addition, these beds form the source material from which much of the overlying glacial deposits were derived.

The Ordovician beds, resting unconformably on the Precambrian basement, have a total thickness of about 1900 feet in the Blue Springs Creek area (Table 1). The lower 1450 feet of the Ordovician beds are

composed of various sedimentary rocks, and are covered by the Upper Ordovician Queenston shale. This is a soft red impervious shale about 460 feet thick. The Ordovician beds outcrop to the east of the basin (Figure 2) and therefore form the source material for much of the glacial deposits in the basin, as the Wisconsinan ice mass that glaciated the basin moved from the southeast (see Section 2.2.1).

The Queenston shale is covered by the Silurian Cataract and Clinton Groups, which together consist of about 85 feet of shales and dolomites (Table 1). None of these beds outcrop in the basin, but they have been observed during the drilling of wells in the area (Figure 3 in pocket). Overlying the Clinton Group is the Middle Silurian Albermarle Group, comprising the Amabel and Guelph Formations. The Amabel Formation has been divided into four members, three of which are present in the Blue Springs Creek basin (Bolton, 1957, Sanford, 1969). The Lions Head Member, which is the lowest member of the Amabel group, occurs in the Bruce Peninsula but is not present in the Blue Springs Creek basin.

The lowest member of the Amabel Formation present in the Blue Springs Creek basin is the Colpoy Bay Member (see Table 1), which is a buff weathered, massive, porous, fine grained to dense, white to blue grey dolomite (Bolton, 1957), lying unconformably on the Clinton Group rocks. The overlying Wiarton Member is a similar dolomite but is light grey with blue grey mottling. Sanford (1969) does not differentiate between the Colpoy Bay and Wiarton members but groups them together as the Wiarton Formation. Bolton (1957), however, places the Colpoy Bay-Wiarton contact at the base of the mottling which is discernible at various exposures in the general area. Both the Wiarton and Colpoy Bay members contain numerous bioherm reefs, which are usually more porous than the surrounding rocks, and are important cave-forming structures. The Wiarton Member was identified in outcrops along the channel of Blue Springs Creek (Map 2). The Colpoy Bay Member, however, was not identified in the basin, but it outcrops or is present just below the streambed along the channel of Blue Springs Creek, as indicated in Figure 3. The overlying Eramosa Member is a bituminous, thin bedded, dark brown, fine grained to sugary dolomite (Bolton, 1957), which outcrops in the extreme western part of the basin (Sanford, 1969). The overlying Guelph dolomite, which is also in the Albermarle Group, is not present in the basin, but outcrops to the west of the area.

Bedrock Topography

Elevation contours of the bedrock surface have been drawn utilizing data from records of water wells drilled in the study area. The locations of these wells are shown on Map 3 and the bedrock surface elevations are shown on Map 4.

The bedrock surface in the Blue Springs Creek basin has a maximum relief of almost 200 feet (60m), which appears more subdued than that of the present land surface. The lowest bedrock elevations are found in the vicinity of Eden Mills where elevations are less than 1050 feet (320m) above sea level; the highest elevations, in excess of 1225 feet (370m) above sea level, occur in the northern portion of the basin.

The most pronounced feature of the bedrock topography is the bedrock valley which is now occupied by the main channel of Blue Springs Creek. This channel was formerly a glacial spillway in front of the Galt and Moffat moraines (Section 2.2.1); further deepening has subsequently taken place through recent erosion of the Blue Springs Creek channel.

TABLE 1. STRATIGRAPHIC COLUMN OF GEOLOGICAL UNITS IN THE BLUE SPRINGS CREEK AREA. AFTER BOLTON (1957) AND SANFORD (1969)

EPOCH	GROUP	FORMATION	MEMBER	CHARACTER OF MATERIAL	APPROXIMATE THICKNESS IN BASIN (Feet) (metres)
Recent		Alluvium		gravel,sand,silt	variable
		Swamp or Marsh		muck,marl,peat	variable
Pleis-tocene		Late Wis-consinan Glacial Drift		silty sand till, kame,esker & out-wash sand & gravel	0-150 (0-50)
Middle Sil-urian	Alber-marle	Guelph (not present in the Blue Springs Creek basin)			0
		Amabel	Eramosa	bituminous brown dolomite	0-30 (0-9)
			Wiarton	massive, light grey porous, mottled, dolomite	30 (10)
			Colpoy Bay	massive, white to blue grey porous, dolomite	50 (15)
			Lions Head (not present in the Blue Springs Creek basin)		0
	Clinton	Reynales		massive light grey dolomite	15* (5)
Lower Sil-urian	Cataract	Cabot Head		grey shale	40* (12)
		Manitou-lin		argillaceous dolomitic limestone	20* (6)
		Whirlpool		massive sandstone	10* (3)
Upper Ordo-vician		Queenston		red shale	460* (140)
Ordo-vician	Various			various	1450* (450)
Precam-brian	Various			various	?

*thickness from gas well, Nassagaweya Tp., Lot 23, Con. III.

Karst Features

Karst landforms consist of features which have been developed predominantly by solution. Typical karst features are developed in limestones, but can also occur in gypsum, rocksalt, dolomite and other water soluble rocks. Dolomite rocks are less soluble in water than limestones (Sweeting, 1973) and therefore have less rapid karst development, as found in the Amabel dolomites in the Blue Springs Creek basin.

One feature of karst areas is the extensive development of underground drainage, which in mature karst leads to the complete lack of perennial surface streams. The initiation of such underground drainage occurs in bedding planes and fractures, which become enlarged due to progressive solution of the rock. The initial development of the solution channels is almost always under phreatic conditions (below the water surface); however, as the openings enlarge, development can continue under phreatic or vadose (free air to water surface) conditions (Ford, 1972). As these conduits enlarge sufficiently, they can eventually form cave structures.

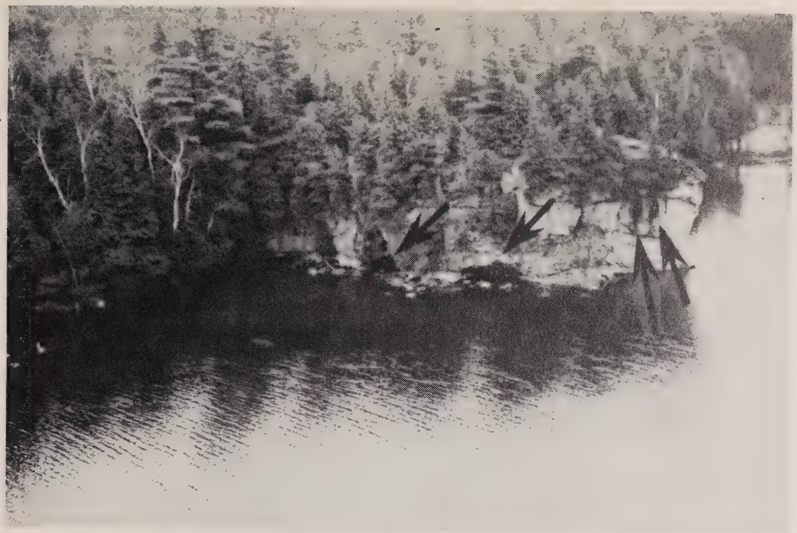
In the Blue Springs Creek basin, no extensive caves have been found; however, small caves do occur nearby at Rockwood, (Photograph 1) one mile to the west of the basin (Map 1). These have been described by Paton (1889), Weber (1960), Kershaw (1973) and Cowell and Gregor (1973).

A number of theories for the origin of the Rockwood caves have been proposed. Weber (1960) thought their formation was the result of an anticline at Rockwood, although this theory has not been supported by other writers. In many other karst areas the solution channels form in the more calcitic and purer members of a limestone/dolomite sequence (Rauch and White, 1970, Sweeting, 1973). Bradley (1968) considered that the Rockwood caves were formed in the bioherms of the Amabel Formation, which he considered were more calcitic. Cowell and Gregor (1973), however, have analyzed the composition of the dolomitic bedrock and the bioherms in the Rockwood gorge for calcite and dolimite by atomic absorption spectrophotometry. They found no significant difference between the compositions of the bioherms and the adjacent bedded dolomites (the mean molar calcium to magnesium ratio of two samples of the bioherms was found to be 1.02, and in three samples of bedded dolomite was 1.05. Ideally, pure dolomite ($CaMg(CO_3)_2$) would have a molar calcium to magnesium ratio of 1.0). The composition of the bioherms therefore, appears to have played no significant part in the cave development.

The Rockwood caves occur near the bioherms of the Amabel Formation (Cowell and Gregor, 1973); however, no caves were found in the bioherm structures themselves, as stated by Karrow (1968) and Weber (1960). The cave structures appear to have developed more readily in the dolomite beds near or at the base of the bioherms, where structuring has disturbed the bedding and increased the porosity of the dolomite.

Cave orientation in the dolomite formation at Rockwood also has been discussed by Weber (1960) and Cowell and Gregor (1973). The latter authors suggest that the orientation is largely controlled by the bioherms. The predominant reef direction (325^{o} from north) (Karrow, 1968) does play a significant part in the orientation of the cave development. This control would also have an influence on the permeability of the dolomite, as conduits would tend to be more open and permeable in the direction of the bioherm reefs.

Other major features that characterize typical karst areas are



Photograph 1: Caves and resurgences found at the conservation area in the vicinity of Rockwood.

Photograph 2: Sinkholes are found at ground surface where dolomite bedrock is exposed or covered by only a thin layer of till.

sinkholes and springs. Both types of features can be observed to a limited extent in the Blue Springs Creek basin.

Sinkholes appear as actual openings in the ground surface, usually connected to vertical solution channels that lead to other underground drainage systems. The sinkholes observed in the study area appear relatively small in size, occurring locally in areas where the dolomite bedrock is exposed or covered by only a thin layer of till (Photograph 2). The 'sinking' phenomena can also be observed along various tributary stream channels where streams actually flow into local depressions and disappear through the soil and overburden materials. This usually occurs in areas where the dolomite bedrock is close to the surface. Surface channels downstream of these sinks are usually very poorly defined as flows very seldom reach the main stream. Several such streams have been identified in the basin (Map 1). Some are intermittent streams, only flowing during wet periods, while others appear perennial in their course.

Two major types of karst springs or resurgences can be found in the area, those where water discharges from vadose channels and the vauclusian type, where the water issues from depth out of phreatic channels (Sweeting, 1973). Photograph 1 shows some springs at Rockwood of the vadose or free flowing type, while the Blue Spring itself, the location of which is shown on Map 1, is of the vauclusian type, the water rising through a mantle of till. Hydrologically, the two types of springs are significantly different (White, 1969). The free flowing springs, as at Rockwood, act like surface streams, exhibiting pronounced floods and droughts; whereas the vauclusian springs are fed from a relatively large ground water reservoir, tending to smooth out any peaky flows. In the Blue Springs Creek basin, the springs shown on Map 1 are largely fed from phreatic channels and so would tend to have less peaky flows than the surface streams.

In typical karst areas the aquifer characteristics, such as transmissivities and storage coefficients, are extremely heterogeneous, with the major portion of the ground water flow taking place along solution channels or caves. In mature karst, the flow of ground water can be represented by flow through a pipe network and not through a granular medium, as is usual for clastic aquifers (White and Longyear, 1962). In attempting to develop piezometric maps for a mature karst area, extreme ground water gradients would likely be found, as adjacent wells may tap water in channels which could have piezometric levels many feet apart. Such a map would therefore contain many discontinuities and many spurious points where the static level would be anomalously low or high. Inspection of the piezometric map (Map 7) of the Blue Springs Creek basin, shows that the piezometric surface is well behaved and that there are very few spurious points. These findings tend to indicate that the karst in Blue Springs Creek is not well developed and that the flow in the dolomite must be taking place through the joints and bedding planes. The ground water flow pattern, therefore, has not been significantly affected by the karst development.

In mature karst areas, the drainage basin (surface and subsurface) is often considerably different in size from the topographic basin, due to water crossing topographic divides in underground solutional channels (White and Schmidt, 1966). In the Blue Springs Creek basin, some small differences have been observed between the topographic and the drainage divides (Map 7), where ground water moves across the topographic divide in the dolomitic joints and bedding planes.

SURFICIAL GEOLOGY

The overburden deposits in the Blue Springs Creek basin consist of glacial drift of Pleistocene age with minor amounts of alluvial and swamp deposits of Recent age. The Pleistocene deposits consist mainly of unstratified till and stratified deposits of kame and outwash materials (Map 2). The thickness of the overburden material (Map 5) ranges from a few feet in the main creek valley, gradually increasing towards the basin divide, to a maximum thickness of almost 150 feet in the northern part of the basin (see also Figure 5).

Glacial History

The glacial features that are present in the Blue Springs Creek basin are the result of the late Wisconsin ice advance, which retreated about 12,000 years ago (Karrow, 1963). The basin was overrun by several pre-Wisconsin ice movements; however, the Wisconsin ice advance has obliterated evidence of these former glaciations in the basin.

During the Wisconsin glaciation, the ice flowed from the present Lake Ontario basin and crossed the Blue Springs Creek area in a west-northwest direction, as is evident from the orientation of drumlins (Chapman and Putnam, 1966). The ice laid down till plains of the Wentworth till, which was derived from the Silurian and Ordovician formations, outcrops of which can presently be found to the east of the area. Drumlins were formed under the ice, and occur as isolated hills on the till plains. As the ice retreated, pauses or possible slight readvances deposited the Paris, and later the Galt and Moffat moraines (Karrow, 1968). Meltwaters flowed down to the southwest, in front of the Paris moraine, and carved out a prominant spillway, part of which is now occupied by the Eramosa River. Meltwaters in front of the Galt moraine flowed southwest from north of Acton and then broke through the Paris moraine at Eden Mills; meltwaters in front of the Moffat moraine flowed across the Galt moraine about two miles to the southwest of Acton (Map 1). The main channel of Blue Springs Creek now occupies these spillways.

As the ice retreated, kames, eskers and outwash plains and terraces were deposited by the sediment laden waters flowing off the ice masses. All these deposits are present in the basin (Map 2) and are discussed in Section 2.2.3.

Wentworth Till

Glacial till can be described as a nonstratified sediment, carried and deposited by glacier ice. Grain sizes in tills vary from clay to boulder size and can be mixed in various proportions. The Blue Springs Creek basin contains the sandy Wentworth till named by Karrow (1963) from its type-area in Wentworth County.

In the southern part of the basin, the Wentworth till appears as a buff coloured, sandy to silty sand till (Karrow, 1968), often containing stones or boulders. These deposits are largely derived from the parent materials of the Lower and Middle Silurian dolomites, shales and sandstones that outcrop to the east of the basin. In the northern part of the basin, the till becomes reddish in colour due to inclusion of the red coloured Ordovician Queenston shale.

The Wentworth till in this area has largely been deposited as the

Paris, Galt and Moffat moraines, which together form a broad morainic belt with a depositional trend from the southwest to the northeast (Map 2). The Paris moraine lies between the Eramosa River and Blue Springs Creek, and the Galt moraine lies to the east of the main Blue Springs Creek channel. The Moffat moraine forms the drainage divide on the extreme eastern part of the basin. Between these moraines, there are glacial spillways which are now occupied by the underfit main channel of Blue Springs Creek. The creek near Eden Mills and the main tributary south of Acton (Map 1), flow in spillways which have cut through the Paris and Galt Moraines, respectively.

Kame, Esker and Outwash Deposits

Kame, esker and outwash deposits are glacial meltwater sediments of stratified sands and gravels. These deposits are often clean washed, although they can include till material; they are usually younger than the tills and are commonly found as surface deposits.

Kames are irregular, hummocky accumulations of sands and gravels containing some till, and are formed from the sediment carried by meltwater on top of the ice. In the Blue Springs Creek basin, kames are abundant in the north and west (Map 2).

Eskers are sinuous ridges of sand and gravel which are formed from sediment laden streams flowing under stagnating ice masses. They tend to be aligned along the direction of ice movement. Several eskers occur in the basin and are shown on Map 2.

Outwash plains and terraces are extensive accumulations of sand and gravel laid down as relatively level deposits whose surfaces are marked by channel scars and kettle holes. In the basin, the outwash sands and gravels occur mainly as terrace deposits along the present course of Blue Springs Creek (Map 2).

Kame, esker and outwash deposits have similar lithologies containing sands and gravels, and are useful sources of building materials. An exposed gravel pit in the northern part of the basin shows a stratified deposit of kame sand and gravel, about 10 feet thick, resting on top of the Wentworth till.

Kame and outwash deposits are usually very permeable and if saturated, can provide high yields to water wells. In the Blue Spring Creek basin, however, the kame and outwash gravels are shallow surficial deposits and are usually above the water table.

Mechanical and Carbonate Analyses of the Pleistocene Deposits

Mechanical analyses of samples of the overburden were carried out to determine the sand, silt and clay content and to compare these to the results obtained by Karrow (1963, 1968) for tills in the Guelph and surrounding areas. Nine samples were taken at different depths from three observation wells that were being drilled in the area. The location of these wells BS-1a, BS-2b and BS-3, are shown on Map 3; the sample depths, well log descriptions, and the results of the mechanical analyses are shown in Table 2. Wells BS-1a and BS-3 were drilled through surface deposits mapped as kame by Karrow, while the map unit at well BS-2b was shown as swamp (Map 2). It was expected that till would be recovered at depth, as kame and marsh are commonly surficial features.

To be consistant with the analyses by Karrow (1968), the mechanical analyses figures given in Table 2 are only those for the fraction smaller

TABLE 2. MECHANICAL ANALYSES AND CARBONATE ANALYSES OF SAMPLES FROM THE BLUE SPRINGS CREEK BASIN

Observation Well No.	Depth (feet)	Material Description from Well log	Mechanical Analysis Clay (%)	Silt (%)	Sand + Gravel (%)	Carbonate Analysis Calcite (%)	Dolomite (%)	Total (%)	Ratio of Calcite to Dolomite
BS-3	10	Brown silty till	2.5	47.4	50.1	14	30	44	0.47
	15	Stony silty till	0	31.0	69.0	11	27	38	0.41
	25	Medium coarse sand and gravel	4.1	22.6	73.3	14	45	59	0.31
	48	Medium coarse sand	3.5	19.8	76.7	16	24	40	0.67
	51	Grey silty till	7.7	30.6	61.7	20	25	45	0.80
BS-2b	10	Buff sandy till and gravel	0	11.6	88.4	3	43	46	0.07
Bs-1a	35	Fine brown silty sand	5.8	63.2	31.0	18	15	33	1.2
	45	Grey silty till	22.0	63.7	14.3	29	8	37	3.6
	85	Stony buff till	6.5	26.9	66.6	17	29	46	0.59
Average of 6 analyses above, recorded as till in the well log.			6.5	35.2	58.4	16	27	44	0.99
Karrow (1968) analysis of Wentworth till near Guelph									
Average of 25 samples			18.0	33.0	49.0	10.6	25.1	35.7	0.63
Minimum values			12.0	22.0	31.0	7.5	13.8	21.6	0.3
Maximum values			33.0	47.0	61.0	19.0	33.4	48.6	1.1

NOTES:

1) The mechanical analysis figures are given for the fraction smaller than 10 mm. This compares with the method used by Karrow (1974, personal communication).
2) Clay-silt boundary is 0.004 mm. Silt-sand boundary is 0.962 mm.
3) Carbonate analyses on fraction passing 200 mesh sieve, using Chittich apparatus.
4) Analyses carried out in the laboratories of the Ministry of Environment.

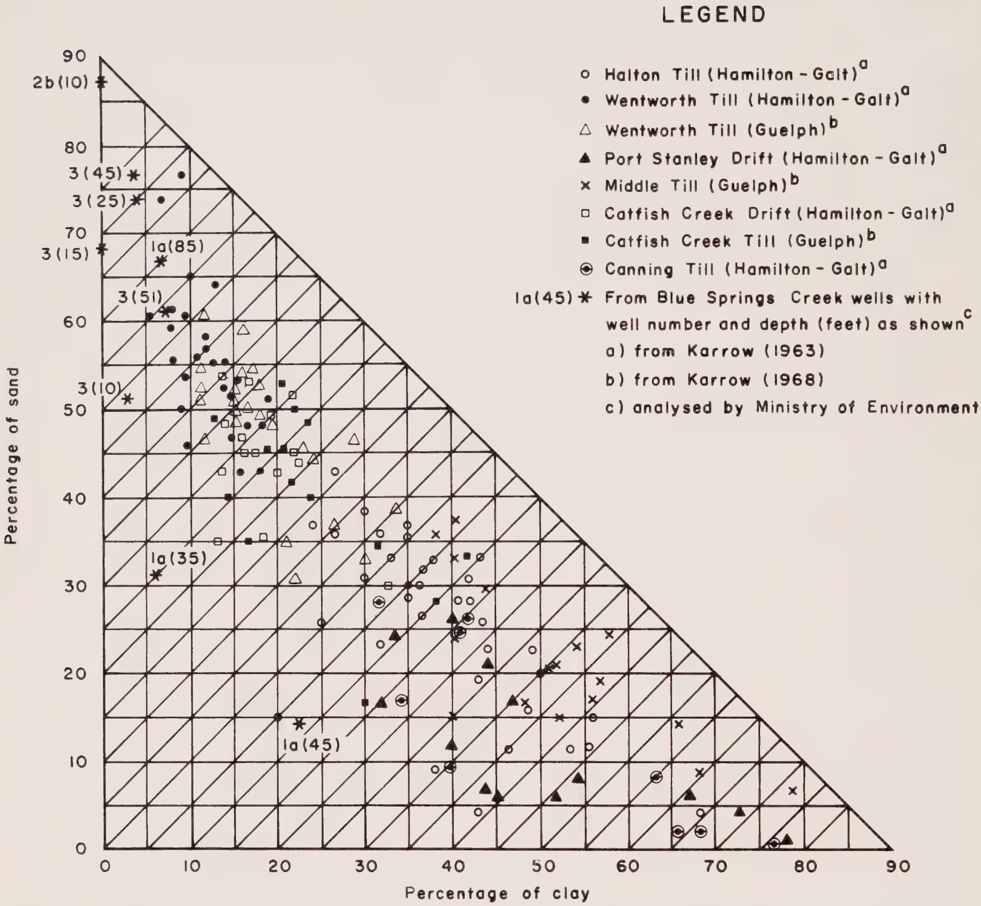


Figure 4. Mechanical analyses of tills in the Blue Springs Creek basin and surrounding areas.

than 10 mm. Karrow (1974, personal communication) had removed all stones larger than about one half inch before performing mechanical analyses on the tills collected in the Guelph area.

The most notable feature of the analyses (Table 2) was the large variability of the clay-silt-sand content of samples found in the Blue Springs Creek basin. To compare these samples to those found by Karrow (1963, 1968) in the Guelph area, and also in the Hamilton and Galt areas to the south of the basin, the figures were plotted on a triangular plot, Figure 4. Most of the samples from Blue Springs Creek basin have less clay than any of the till samples obtained by Karrow, and only three samples have compositions similar to those of the tills in the Guelph or Hamilton and Galt areas. This shows that most of the samples tested in the Blue Springs Creek basin were from thick kame deposits and were not till deposits.

Carbonate analyses were also carried out on the samples from the basin and the results are recorded in Table 2. Similar to the mechanical analyses, there were larger variations in the carbonate figures from the basin samples than in those obtained by Karrow (1968) from the tills in the Guelph area. The carbonate analysis figures confirm that the samples tested were not till.

RECENT GEOLOGY

Recent deposition can be found as alluvial and swamp deposits. The alluvial deposits, composed of gravel, sand, silt and clay, are not widespread in the basin, occurring only in the Blue Springs Creek valley near Eden Mills (Map 2). Bogs and swamps are common in the flat, poorly drained areas of the basin and in areas where ground water discharge is taking place. Swamps are often covered with shrubs and contain peat, muck and marl.

Soils

The type of soil formed in an area is related to the type of parent material, the drainage and the topography of the area. The soils in the Blue Springs Creek basin are derived from the surficial glacial material and from the dolomite bedrock exposed near the creek. The soils in the area have been mapped by Hoffman et al (1963) and by Gillespie et al (1971), and are shown on Map 6.

The predominant soil is the Dumfries loam, which is derived from the Paris and Galt moraine material, and is usually stony and unproductive. On the till plains where the topography is more subdued, the Guelph loams have developed. These are better soils but can also be stoney or may have high relief, hindering crop production. The esker and kame deposits and the bedrock exposures near the creek develop a variety of poorer soils (Map 6).

The soils map can be used to delineate moist areas where the water table is near ground level. The Dumfries, Killean, Lily, Mesisol and marsh soils are a progressively wetter succession of soil types.

The Mesisol and marsh areas (Map 6) occur in several places in the Galt moraine and can be considered as areas where evapotranspiration is high, even during climatically dry periods in the summer. Marsh and Mesisol areas are also considered as possible ground water discharge areas. Conversely, well drained soils such as the Dumfries, or thin soils such as the Donnybrook Loam, are indicative of areas of relatively low evapotranspiration during dry periods.

HYDROGEOLOGY

This chapter presents an analysis of the hydrogeologic factors affecting the behaviour of the ground water regime in the basin. The data were compiled mainly from water well and observation well records maintained by the Ministry of the Environment. Supplementary data, such as streamflow and precipitation records, used in arriving at estimates of ground water discharges, were obtained from station networks maintained by the Ministry, the University of Guelph and Environment Canada (Water Survey of Canada and Atmospheric Environment Service).

DELINEATION OF HYDROGEOLOGIC UNITS AND PRINCIPAL AQUIFERS

On the basis of well data, geologic and hydrogeologic cross-sections were constructed approximately perpendicular to the axis of the main Blue Springs Creek, as shown in Figure 5. Based on these cross-sections, three hydrogeologic units were identified, as shown in Table 3.

TABLE 3. HYDROGEOLOGIC UNITS AND PRINCIPAL AQUIFERS IN THE BLUE SPRINGS CREEK BASIN

<table>
<tr><th>Unit</th><th>Hydrogeologic Unit</th><th>Principal Aquifer</th></tr>
<tr><td>1</td><td>kame, outwash and alluvial sand and gravel</td><td rowspan="2">sandy till reservoir</td></tr>
<tr><td>2</td><td>sandy till</td></tr>
<tr><td>3</td><td>bedrock</td><td>bedrock aquifer</td></tr>
</table>

The kame, outwash and alluvial sand and gravel unit occurs generally as lenses or pockets overlying the sandy till unit; however, it forms only minor ground water aquifers in the Blue Springs Creek basin.

The role of the sandy till unit is that of an aquitard. Sand and gravel lenses occur in the sandy till unit, but for the most part are not interconnected and have little influence on the hydrogeologic behaviour of this unit. As vertical leakage from unit one to unit two is dictated by the permeability of the till, the two units are grouped together into one principal aquifer referred to as the sandy till reservoir.

Underlying the sandy till reservoir is the bedrock unit which constitutes the major aquifer in the study area. This unit, referred to as the bedrock aquifer, is the major source of ground water supply, and discharges significant amounts of ground water to the main Blue Springs Creek.

COEFFICIENTS OF TRANSMISSIBILITY AND STORAGE

In ground water investigations, the coefficient of transmissibility (called the transmissivity in this report) and the storage coefficient,

are important parameters that control the rate and quantity of the ground water flow in a basin. The transmissivity in the Blue Springs Creek basin was estimated using various techniques of analysis. These were:

1) Pumping tests on wells (Section 3.2.1)
2) Specific capacity of wells (Section 3.2.2)
3) Ground water flow rates (Section 3.2.3)
4) Ground water modelling (Section 4.4).

The storage coefficient was estimated by using the following methods:

1) Pumping tests on wells (Section 3.2.1)
2) Ground water recharge and discharge rates (Section 3.3.1)
3) Ground water modelling (Section 4.5).

Hydraulic Properties of the Aquifers Based on Pumping Test Data

The transmissivities and storage coefficients of the aquifers were estimated from two pumping tests carried out in the basin. In addition, the results of two tests carried out near the village of Rockwood, which is just outside of the basin, were included.

There are a number of ways of calculating aquifer constants from pumping test data. For this study, three methods of analysis were used, all based on the Jacob well equation (Cooper and Jacobs, 1946). The methods used were:

1) Measurement of drawdown in an observation well during pumping,
2) Measurement of drawdown of the pumped well during recovery,
3) Drawdown-distance method, using the drawdowns in observation and pumped wells at the end of the pumping period.

The derivation of the equations and the analysis method are described in Todd (1959) and Wisler and Brater (1959).

Wells BS-2a and BS-4a were test-pumped in 1966, after being drilled for use as observation wells. Wells TW1-67 and TW2-67 were drilled and test-pumped in 1967 during a Ministry ground water survey for the Village of Rockwood (Sobanski, 1968). The location of these wells are shown on Map 3 and the results of the pumping tests are given in Table 4.

At TW1-67, there was a marked discontinuity in the drawdown-time plot after 20 minutes. This was interpreted as being due to a region of high transmissivity in the vicinity of the well surrounded by an area of lower transmissivity (Sobanski, 1968). The transmissivities found for the bedrock aquifer ranged from 1000 to 12000 IGPD/ft, with a mean of approximately 4200 IGPD/ft. As indicated by the range of the transmissivity figures obtained during the tests, the bedrock aquifer is considered to be quite heterogeneous.

Only one pumping test was carried out in the overburden aquifer. The transmissivity found from this test was 6800 IGPD/ft (Table 4).

The coefficient of storage around the pumped wells was also estimated (Table 4) and averaged 2×10^{-5}. This figure is slightly lower than that which is commonly encountered in confined aquifers, where the coefficient of storage is usually between 5×10^{-5} and 500×10^{-5} (Todd, 1959). The coefficient of storage is directly related to the thickness and porosity of the aquifer (Murray, 1970); a low storage figure usually indicates a low porosity and/or a thin aquifer. As the Amabel Formation is fairly thick (approximately 180 ft. near Rockwood) and has a low coefficient of storage, it can be considered to have a relatively low

TABLE 4. ESTIMATED TRANSMISSIVITIES AND STORAGE COEFFICIENTS DETERMINED FROM PUMPING TESTS

Well No*	Aquifer and Depth (ft)	Pumping Rate (IGPM)	Distance to Observation Well (ft)	Draw-down (ft)	Method**	Time of Pumping or Recovery (hours)	Estimated Transmissivity (IGPD/ft)	Estimated Storage Coefficient (Dimensionless)
BS2a	Dolomite (45-50)	4 and 6	237	13	D-T	4	4500	-
			237		D-D	4	1000	-
BS4a	Sand & Gravel (26-29)	33-43	9.5	16.5	D-T	6 1/3	6800	-
TW1-67***	Dolomite (22.5-195)	200	700	24.1	D-T	24	6000	1.5×10^{-5}
			-		D-T-R	8	12000 up to 20 minutes 4500 after 20 minutes	- - - -
TW2-67***	Dolomite (15-210)	50	350	41.	D-T	10 2/3	1000	2.5×10^{-5}
			350		D-D	10 2/3	2900	-
			-		D-T-R	2	1900	-

* Location shown on Map 1

** D-T Drawdown-Time for Pumping
D-T-R Drawdown-Time for Recovery
D-D Drawdown-Distance

*** After Sobanski (1968)

porosity. This low porosity is further supported by the low value of absorption found for the Amabel dolomite. The absorption is the amount of water that a piece of dry rock can absorb, if immersed in water. The absorption was measured at 0.93 per cent for the Amabel near Adamsville (Hewitt, 1964b). The low porosity and low coefficient of storage support the idea of the flow being confined to channels along joints and bedding planes, rendering the effective thickness of the aquifer considerably smaller than the total measured thickness.

Evaluation of Transmissivity from Specific Capacity Data of Wells

Transmissivity values were estimated from well records to provide an evaluation of the aquifer characteristics throughout the Blue Springs Creek basin. The bedrock and overburden aquifers were compared, and the transmissivity in the bedrock aquifer was further related to the aquifer depths and piezometric surfaces.

The specific capacity of a well is the rate of discharge of water from the well (expressed in gallons per minute), divided by the drawdown (in feet). The specific capacity depends on the aquifer characteristics, the well diameter, the pumping time, the screening in the well and a number of other factors. However, a high specific capacity of a well generally indicates a high transmissivity of the aquifer.

The relation between the specific capacity of a well and the transmissivity has been studied by Theis et al (1954). They calculated the theoretical specific capacities of wells in relation to the transmissivity and coefficient of storage of the aquifer, the pumping period and the well diameter, assuming that the well was fully penetrating and was fully developed. From the data presented in Theis et al (1963) the relation between the transmissivity and the specific capacity is shown in Figure 6, for a well diameter of 0.5 feet, storage coefficients of 0.1 and 2×10^{-5} and pumping periods of 4 hours and 1 day.

The transmissivity values near each well were calculated from the specific capacity data by using the relationship shown in Figure 6. Most of the wells were approximately 6 inches in diameter; if the well was considerably larger or smaller, the relationships given in Theis et al (1963) were used to estimate the transmissivities. Most of the bedrock wells were assumed to be under confined conditions, with the storage coefficient taken as 2×10^{-5}. Most of the overburden wells were assumed to be under water table conditions, with the storage coefficient taken as 0.1. Wells having a low specific capacity (less than 0.1 IGPD/ft) and a short pumping time (less than 2 hours) were excluded from the analysis, as the yield from such wells is largely derived from water held in storage inside the well itself and not from the aquifer. Due to the assumptions inherent in this method, the transmissivity values obtained from this analysis should be regarded as approximations.

In order to determine the median values and the range of values obtained for the transmissivities, a statistical analysis (after Siddiqui and Parizek, 1971) was applied to various categories of wells. The transmissivity values in each category were listed in ascending order of magnitude and assigned probabilities according to the relationship:

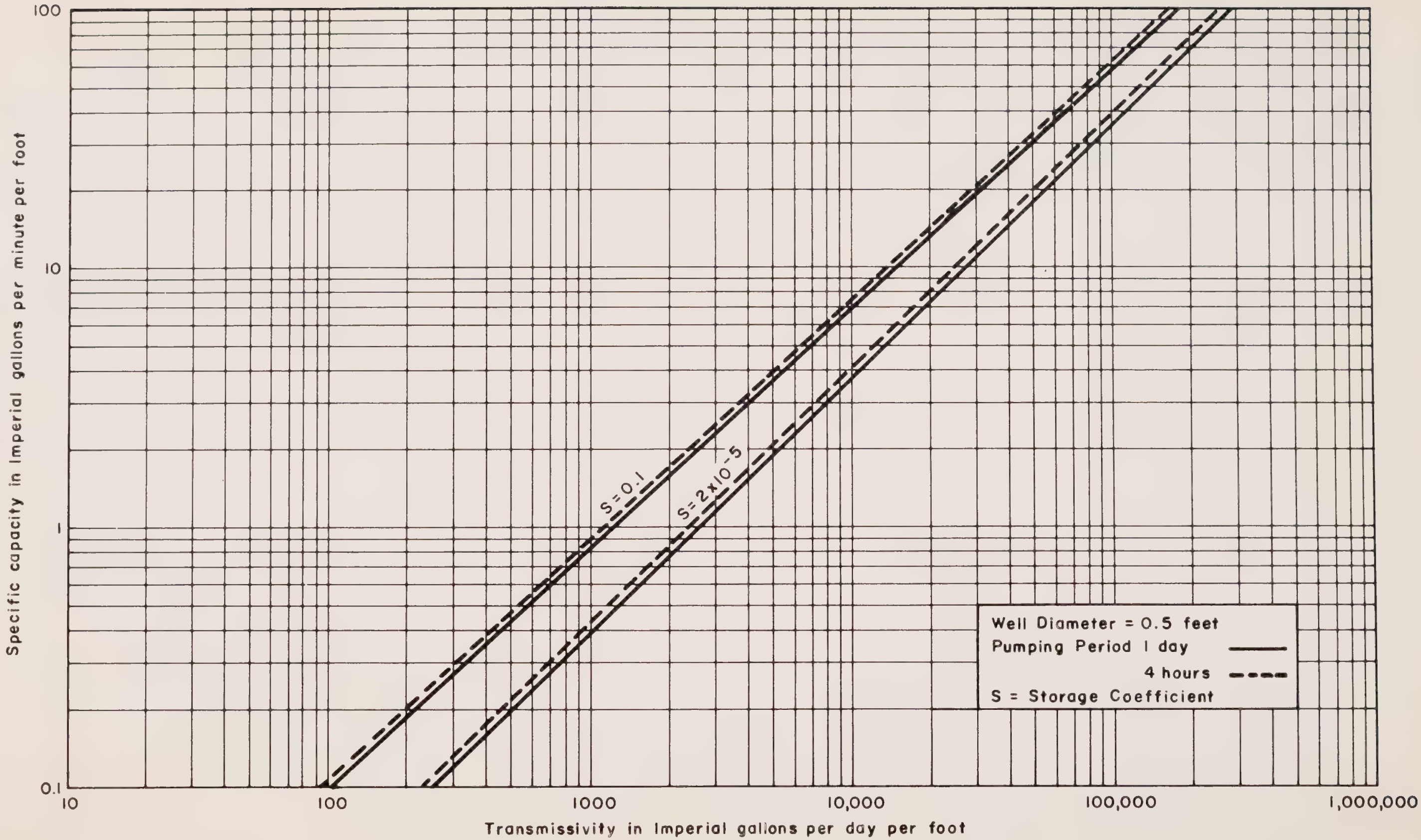


Figure 6. Relationship between specific capacity and transmissivity.

$$F = \frac{100m}{(n+1)} \qquad (1)$$

where F = percentage of wells where transmissivities are less than the transmissivity of a well of serial number m,

m = serial number of well arranged in ascending order of transmissivities,

n = total number of wells.

The transmissivity values were then plotted against percentage of wells on logarithmic probability paper. This technique implies that a straight line can be fitted to the plotted points if the observations are a random sample from a log-normal distribution.

The analysis was carried out for wells penetrating the overburden, the bedrock, and for bedrock wells which penetrated aquifers at various depths and had different static water levels.

Overburden Wells .... In order to estimate overburden aquifer characteristics in the region, data for eight overburden wells in the basin and 49 overburden wells from the immediate surrounding areas were analysed (see Map 3).

The logarithmic probability plot for the overburden wells is shown in Figure 7, and the number of wells, 10 percentile, median and 90 percentile transmissivity figures are shown in Table 5. The 10 and 90 percentile figures are the transmissivities not exceeded by 10% and 90% of wells, respectively. These figures give a measure of the dispersion of the transmissivity values; a large difference between the 10 and 90 percentiles indicates a large spread and a high standard deviation. The standard deviation can be given by (Brookes and Dick, 1951):

$$\sigma = \frac{X_{(90)} - X_{(10)}}{2.57} \qquad (2)$$

where $X_{(90)}$ and $X_{(10)}$ are the 90 and 10 percentile readings, respectively. As the transmissivities are assumed to be log-normally distributed, the standard deviations which are recorded in Table 5, are of the log-transformed data. The log transform of the median figures are also recorded in this table for use in statistical analyses (Section 3.2.2.3).

From Figure 7, it can be seen that the points plot on a straight line, supporting the hypothesis that the transmissivity values are distributed log-normally. The wide variation in the transmissivity figures can be seen from the percentile figures in Table 5, indicating the heterogeneity of the overburden aquifer.

Bedrock wells .... In the basin, 186 bedrock wells were recorded, 99 of which were selected at random and used for transmissivity analysis. The transmissivity probability plot for these 99 wells is shown in Figure 8 and the characteristics of the plot are recorded in Table 5. It can be seen from Figure 8 that the lower transmissivity values plot on a straight line, whereas the higher values deviate from this line. There are two possible explanations for this deviation.

The specific capacity of a well is the pumping rate divided by the drawdown. At high transmissivities, the drawdown is small and is

TABLE 5. TRANSMISSIVITY VALUES FOR WELLS IN DIFFERENT AQUIFERS IN THE BLUE SPRINGS CREEK BASIN

		Transmissivity Values (IGPD/ft)			Log-transformed data	
Classification	Number of wells	10 percentile	Median	90 percentile	mean	standard deviation
Overburden	49	300	3500	42000	3.545	0.834
Bedrock	99	610	3200	17000	3.506	0.562
Bedrock Pumping Tests (Table 4)	8	750	3100	13000	3.491	0.482
Bedrock grouped by depth to static level						
(feet)						
<14	27	400	4200	16000	3.624	0.625
15-24	26	630	3200	45000	3.506	0.723
25-39	23	500	3000	30000	3.478	0.692
>40	23	900	2750	12000	3.440	0.435
Bedrock grouped by depth to water found						
(feet)						
<44	22	370	3900	50000	3.591	0.829
45-69	30	490	3400	37000	3.533	0.731
70-89	22	1050	3100	8800	3.492	0.361
>90	25	600	3200	11500	3.506	0.497

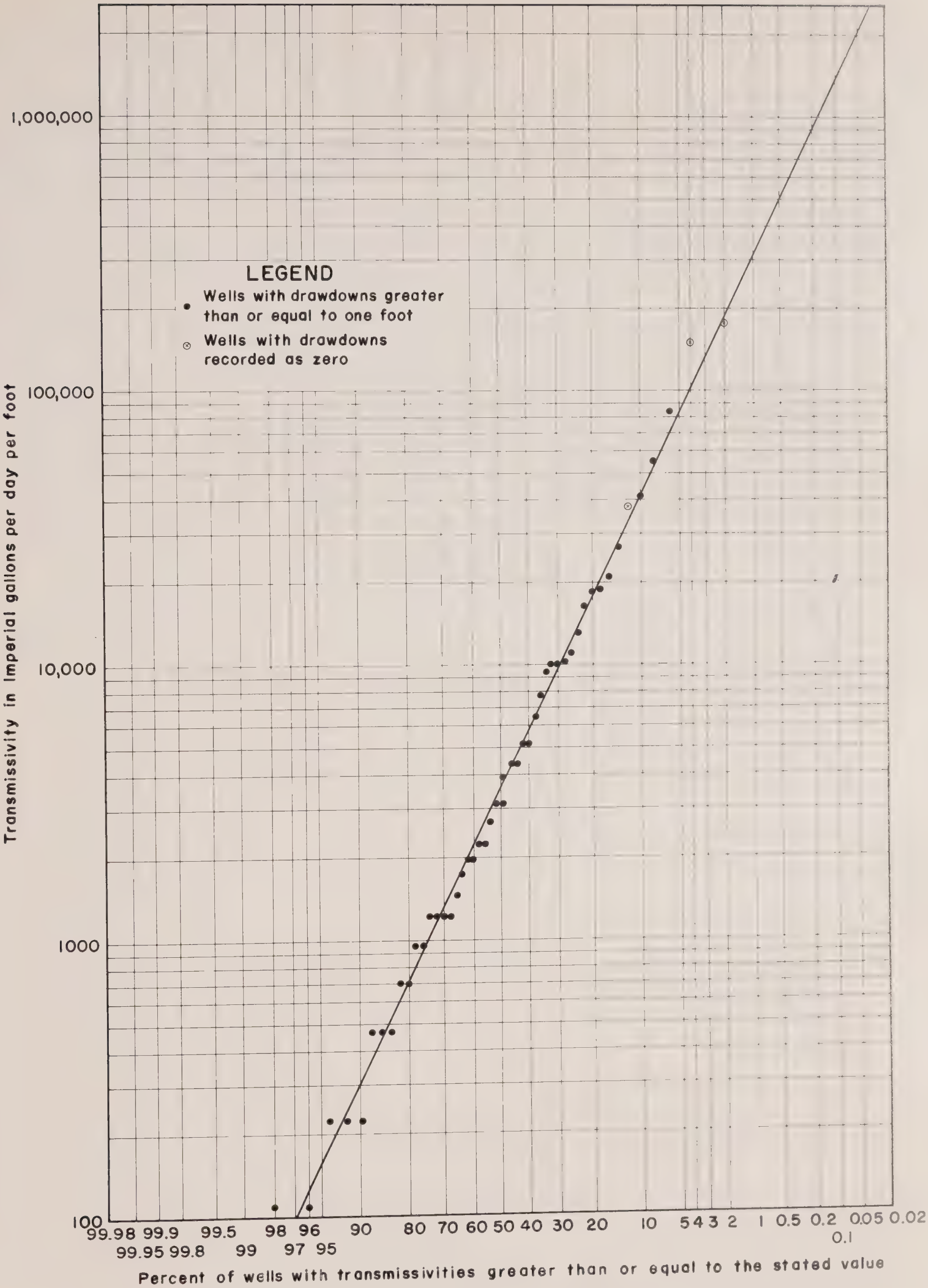


Figure 7. Transmissivity - probability graph for overburden wells.

sometimes reported as zero. In these cases, the specific capacity was calculated by assuming the drawdown to be 0.5 feet, introducing a possible error. The points where the drawdown was recorded as zero are marked on Figure 8; it can be seen that these include most of the high transmissivity figures.

The other possible explanation is that some of the high transmissivity wells are intrinsically different. The majority of the wells are tapping the fissured dolomite; however, a few wells could penetrate solutional channels in the dolomite, which would considerably elevate the transmissivity values.

Some of the overburden wells showed zero drawdown during short pumping tests (Figure 7) and yet did not display such a deviation at the higher transmissivity values. For this reason, the bedrock wells are probably from two different populations, with some of the wells tapping solutional channels or highly fissured and therefore highly permeable portions of the dolomite.

The eight bedrock transmissivity values obtained from the pumping tests (Section 3.2.1) were also plotted on Figure 8. They plot very close to the line obtained from the specific capacity of the wells. This shows that the transmissivity values obtained from the specific capacity data are comparable to the transmissivities obtained from the pumping test data.

Comparison of overburden and bedrock transmissivity values .... From Table 5 it can be seen that the median transmissivity values for the bedrock and the overburden wells are similar at 3200 and 3500 IGPD/ft, respectively. A Student 't' test (Gregory, 1963) on the log-transformed data showed that there was no significant difference between these values at the one per cent level.

The standard deviation of the overburden wells, however, is larger than the standard deviation of the bedrock wells (Table 5). The 'F' test (Hays, 1973) showed that the difference between the standard deviations of the bedrock and overburden wells is significant at the one per cent level.

The median values of transmissivities of the two aquifers are similar; the overburden aquifer, however, has more variability than the bedrock aquifer.

Bedrock well analysis .... Further analysis was performed on the bedrock wells to determine if the depth where water was found or the depth to the static water level had any relation to well yield. The bedrock wells were split into four categories according to the depths to static water level and depths to water found and plotted in figures 9 and 10. The characteristics of the curves are recorded in Table 5. It was found that the points in figures 9 and 10 did not lie on straight lines, but due to the small sample size (generally about 25), it was considered likely that the transmissivity values were log-normally distributed, and that the deviations seen were random. Standard significance tests were carried out on the assumption that the values were log-normally distributed.

For the depths to static water level, the median transmissivity values in each category in Table 5 appear to increase as the water level becomes shallower, as was found in Pennsylvanian rocks by Siddiqui and Parizek (1971). However, these increases were found to be not significant at the 1 per cent level. In addition, the standard deviations of the figures are not significantly different at the one per cent level between any two categories.

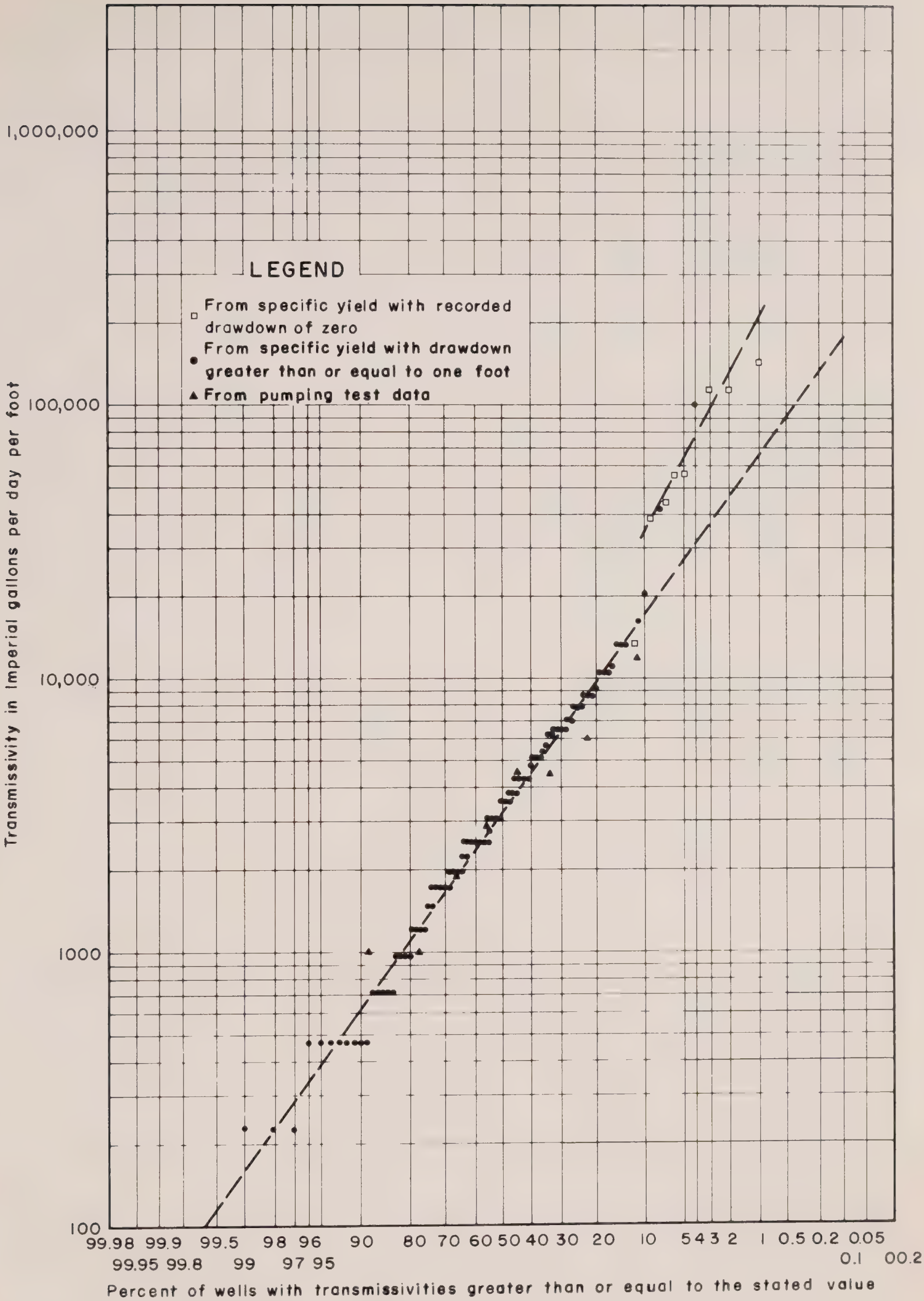


Figure 8. Transmissivity — probability graph for bedrock wells.

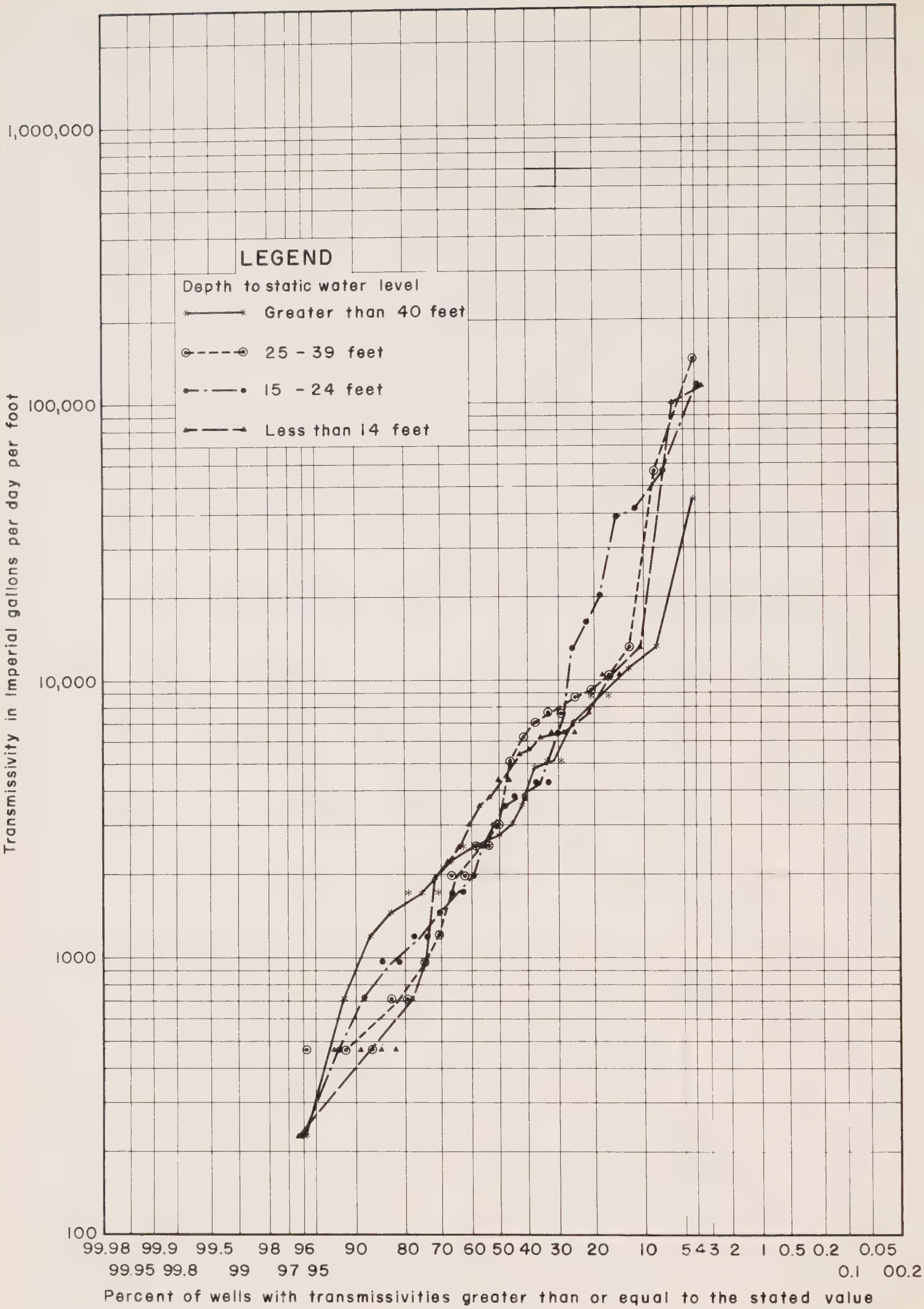


Figure 9. Transmissivity – probability graph for bedrock wells grouped according to static water level.

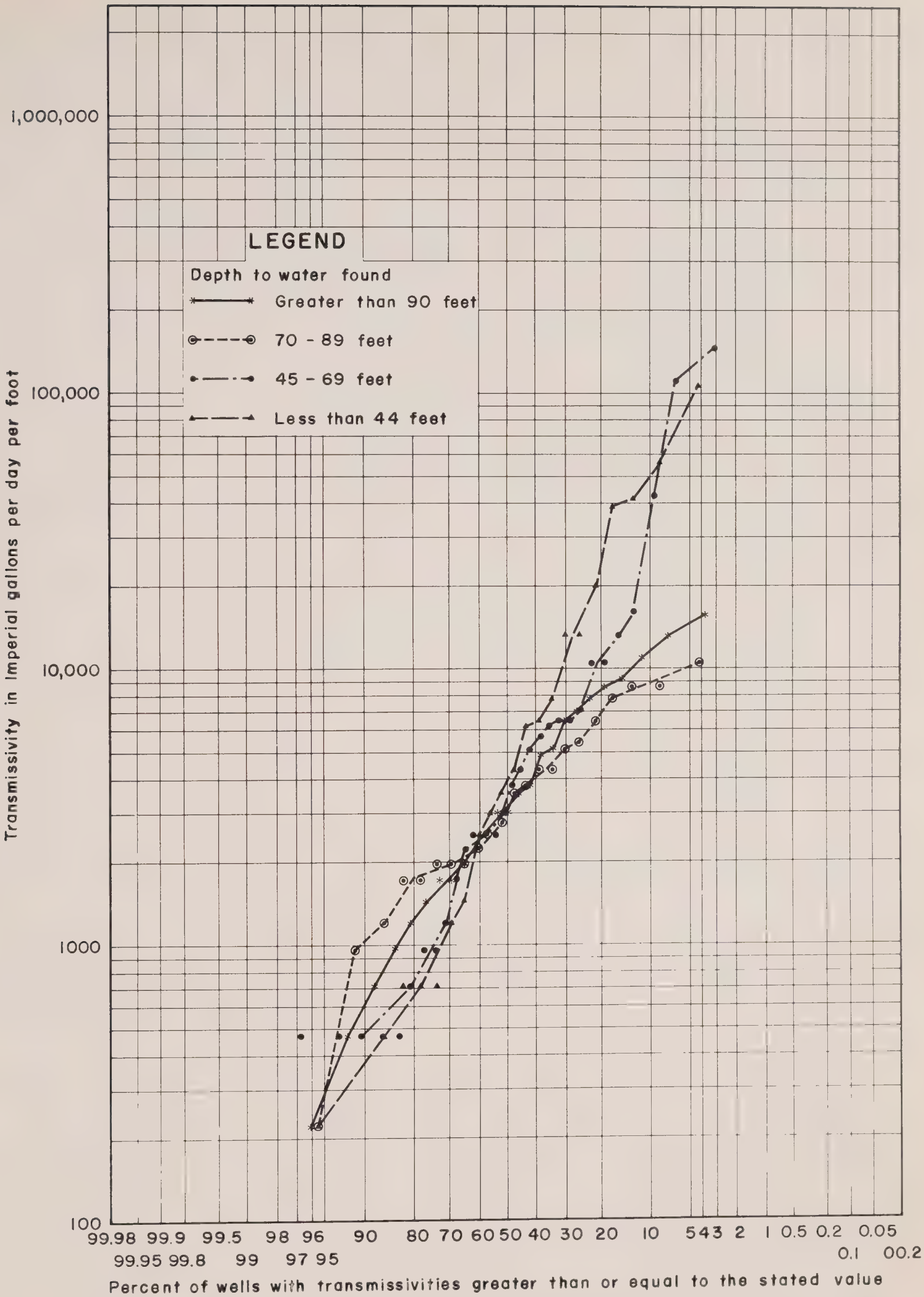


Figure 10. Transmissivity – probability graph for bedrock wells grouped according to depth to water found.

The values of depth to water found are plotted in Figure 10 and the characteristics of these points are also listed in Table 5. This analysis differentiates the wells according to the depths to the water-yielding zones in the bedrock. No significant differences were found between the median figures; however, a significant difference was found between the variability in the categories <69 feet and those >70 feet to water found. Thus, the wells tapping shallow aquifers do show a larger variation than wells tapping deep aquifers. This finding supports the theory of shallow development of solutional channels (Swinnerton, 1932), where the range of transmissivity values would be expected to be larger than in deeper portions of the rock with less well developed solutional channels. It is not known, however, why the median transmissivity values in this basin are not significantly different for different depths to water found.

Computation of Transmissivity Values Using Ground Water Flow Rate

The mean regional transmissivity figure can be estimated by considering the ground water flow rate into the streams. In any aquifer, the ground water flow is governed by Darcy's Law, which for discharge to streams can be stated as:

$$Q = 2TIL \qquad (3)$$

where Q is the mean gain from ground water flow from both sides of the stream along a reach of length L,

T is the mean regional transmissivity,

and I is the mean hydraulic gradient towards the stream.

The estimation of the transmissivity was carried out for the section of stream between the federal gauge and the Boy Scouts Camp gauge, also shown in Figure 15. This reach is near the centre of the basin, is relatively straight and draws its baseflow from the bedrock aquifer.

The stream length is 2.9 miles and the mean ground water gradient, as measured from Map 7, is 84 ft/mile on the west side and 112 ft/mile on the east side of the creek. The mean baseflow contribution from 1968 to 1971 to the creek between the Boy Scouts Camp and the federal gauge, was 3.9 MIGPD.

Using the mean ground water gradient and the measured values for the length and baseflow with Equation 3, the mean transmissivity ($\overline{T}$) was found to be 7000 IGPD/ft. This value is an estimate of the mean transmissivity of the bedrock aquifer adjacent to the creek between the Boy Scouts Camp and the federal gauge. It can be compared to the median values determined in the previous sections (3.2.1 and 3.2.2) on well analysis, if the distribution of the transmissivity values is considered.

The transmissivity values of the bedrock aquifer in this basin were found to be approximately log-normally distributed (Section 3.2.2). It can be shown (Kendall, 1966), that the arithmetic mean of the transmissivity is related to the median by the relationship:

$$\bar{T} = \exp\left(\bar{X} + \sigma_x^2/2\right) \qquad (4)$$

where $X = \ln(T)$ (5)

$\bar{T}$ is the arithmetic mean of the transmissivity values

$\bar{X}$ is the median transmissivity value

and σ_x is the standard deviation of the log-transformed data

Using the values for $\bar{X}$ and σ_x derived from Table 5, the mean transmissivity of the bedrock aquifer can be estimated, based on the specific capacity of the wells as being 7400 IGPD/ft. Thus the arithmetic mean transmissivity found from well analysis agrees very closely with the transmissivity derived from the ground water flow rate.

GROUND WATER MOVEMENT

In most temperate basins containing clastic rocks, the regional ground water flow is from the high points in the basin towards the lower streams or springs. The streams are usually effluent (Meinzer, 1942) and are continuously being recharged by ground water. In any karst area, however, the situation can be complicated by the presence of well developed solutional channels which can transmit water across topographic divides (for example White and Schmidt (1966)) or interfere with normal flow of ground water into the streams.

A piezometric contour map was prepared for the Blue Springs Creek area, using bedrock water well information, as shown on Map 7. From this map the approximate location of the ground water divide was found to be not coincident with the topographic divide. The deviation occurs in the southeastern part of the basin where ground water is lost from the topographic basin, and in the northwestern part of the basin where ground water flows into the basin. The ground water basin has an area of 26.4 square miles (68.4 km^2), which is 13 per cent smaller than the topographic basin area (30.2 square miles or 78.2 km^2). Because of the smaller ground water basin area, the baseflows in the stream will be smaller than those calculated using the topographic basin area.

Monthly Percolation and Ground Water Recharge

Monthly amounts of percolation and ground water recharge in the Blue Springs Creek basin were estimated by determining the soil water and the ground water budgets. The hydrologic factors which may affect the percolation and ground water recharge include precipitation, evaporation, snow depth, overland flow, ground water flow, ground water storage and soil storage. The interaction of these factors can be illustrated in the simplified conceptual model shown in Figure 11. In this figure, there are four storage zones (the snowpack, the soil, the lower soil and the ground water), where the conservation of mass equation can be applied:

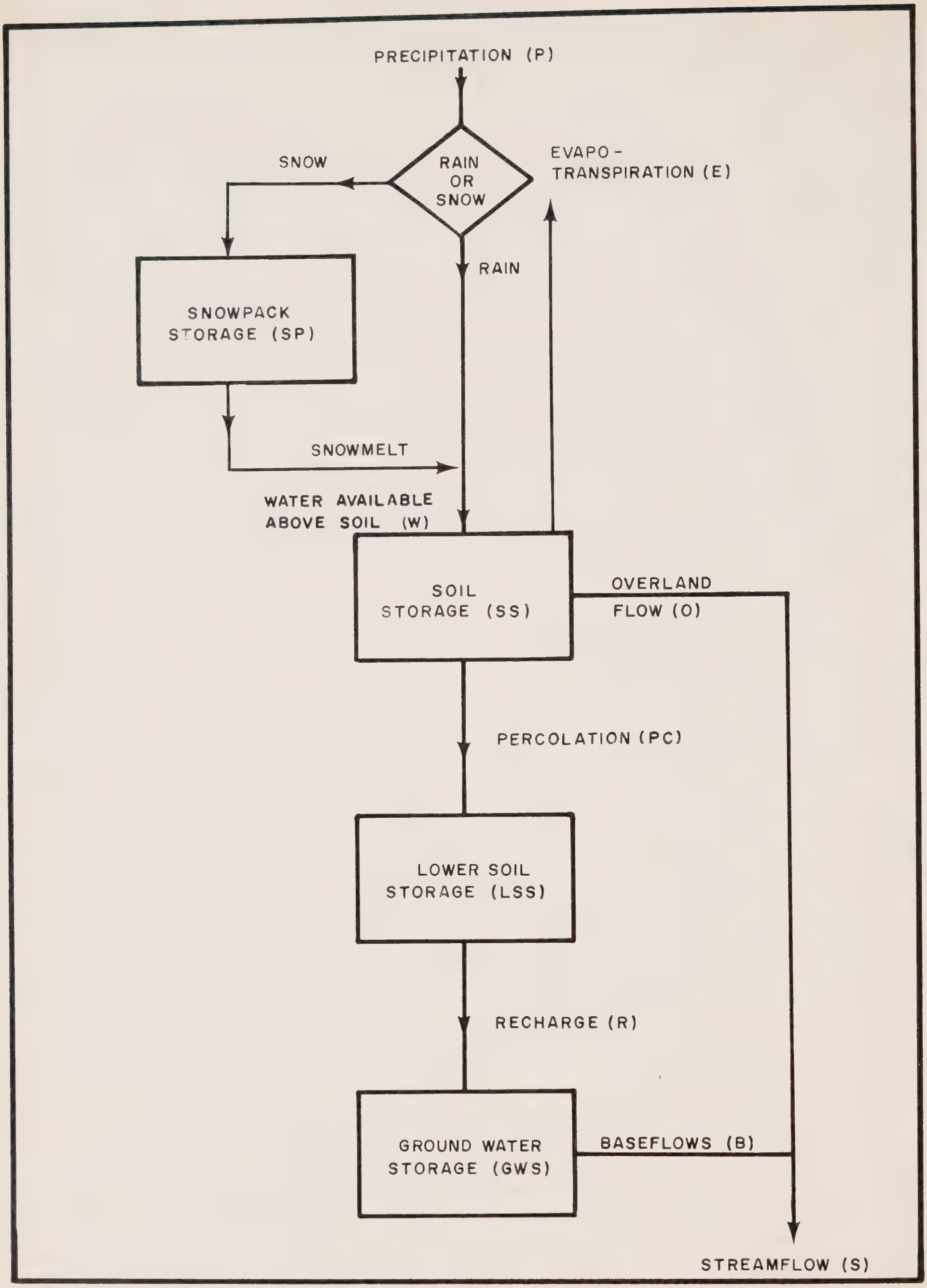


Figure 11. Conceptual model used to estimate percolation and ground water recharge.

in the snowpack

$$P = W + \Delta SP; \qquad (6)$$

in the soil

$$W = PC + O + E + \Delta SS; \qquad (7)$$

in the lower soil,

$$PC\,\cancel{I} = R + \Delta LSS; \qquad (8)$$

and in the ground water

$$R = B + \Delta GWS \qquad (9)$$

where P is the precipitation,
W is the water available above the soil,
ΔSP is the change in snowpack storage,
PC is the percolation below the soil zone,
O is the overland flow,
E is the evapotranspiration,
ΔSS is the change in soil storage,
ΔLSS is the change in lower soil storage,
R is the ground water recharge,
B is the baseflow,
and ΔGWS is the change in ground water storage.

The evapotranspiration in Equation 7 implicity includes any snowpack and ground water evaporation.

The mean monthly recharge was estimated for the years 1966 to 1972, by solving equations 6 to 9. Much of the data in these equations are available from existing records; the precipitation was taken from the meteorological records, the evapotranspiration was calculated by the Thornthwaite (1948) method, the snowpack storage was taken as one tenth of the snowpack depth at the Blue Springs Creek basin, and the streamflow at the federal station was separated into the baseflow and overland flow components. These data are listed in Appendix 1. Using these data, the only parameters in equations 6 to 9 that are unknown are the soil storage, the percolation, the ground water storage, the lower soil storage and the recharge.

The soil storage was estimated for each month by making the following assumptions concerning the percolation and maximum soil storage levels. The soil was allocated a maximum storage of the field capacity, and any excess water available was allowed to percolate. If the soil was below field capacity, the percolation was considered to be zero.

Twenty measurements were made of the moisture storage in the top four feet of the soil zone at three sites during 1969 in the Blue Springs Creek basin. The measured values range from about 10 inches in the spring to three inches in the late summer. It was found that these soil moisture measurements were closely modelled if the field capacity was set at 10 inches. Using this figure for the field capacity, the monthly percolations were determined from Equation 7. The data are shown in Appendix 1 and the mean annual values are shown in Table 6. The monthly percolations are plotted in Figure 12.

TABLE 6. WATER BUDGET IN THE BLUE SPRINGS CREEK BASIN; ANNUAL SUMMARY, 1966-1972

Water Year	P	E	S	B	O	MWH	GWS	SS	PC	R	LSS
(October 31, 1966)						24.14	10.00	10.0			10.0
66-67	38.89	19.26	14.60	9.09	5.51	21.02	11.49	8.14	11.98	10.56	11.42
67-68	37.55	20.18	16.41	10.58	5.83	20.44	11.77	10.0	9.67	10.83	10.26
68-69	31.23	20.66	17.86	12.45	5.41	22.32	10.87	3.85	11.31	11.52	10.05
69-70	32.72	20.66	10.53	7.08	3.45	21.94	11.05	8.39	4.07	7.20	6.92
70-71	36.58	20.40	16.29	10.04	6.25	20.88	11.55	8.18	10.14	10.50	6.56
71-72	35.90	20.95	16.02	9.71	6.31	21.34	11.34	6.16	10.66	9.42	7.80
Mean	35.48	20.35	15.29	9.83	5.46				9.64	10.01	

P	Annual precipitation, in inches,
E	Annual evapotranspiration, in inches.
S	Annual streamflow, in inches.
B	Annual baseflow, in inches.
O	Annual overload flow, in inches.
MWH	Mean water level in the six observation wells on October 31, in feet below ground surface.
GWS	Relative ground water storage on October 31, in inches: Initial level assumed.
SS	Soil storage level on October 31, in inches: Initial level assumed.
PC	Annual percolation in inches.
R	Annual ground water recharge in inches.
LSS	Relative lower soil storage level in inches: Initial level assumed.

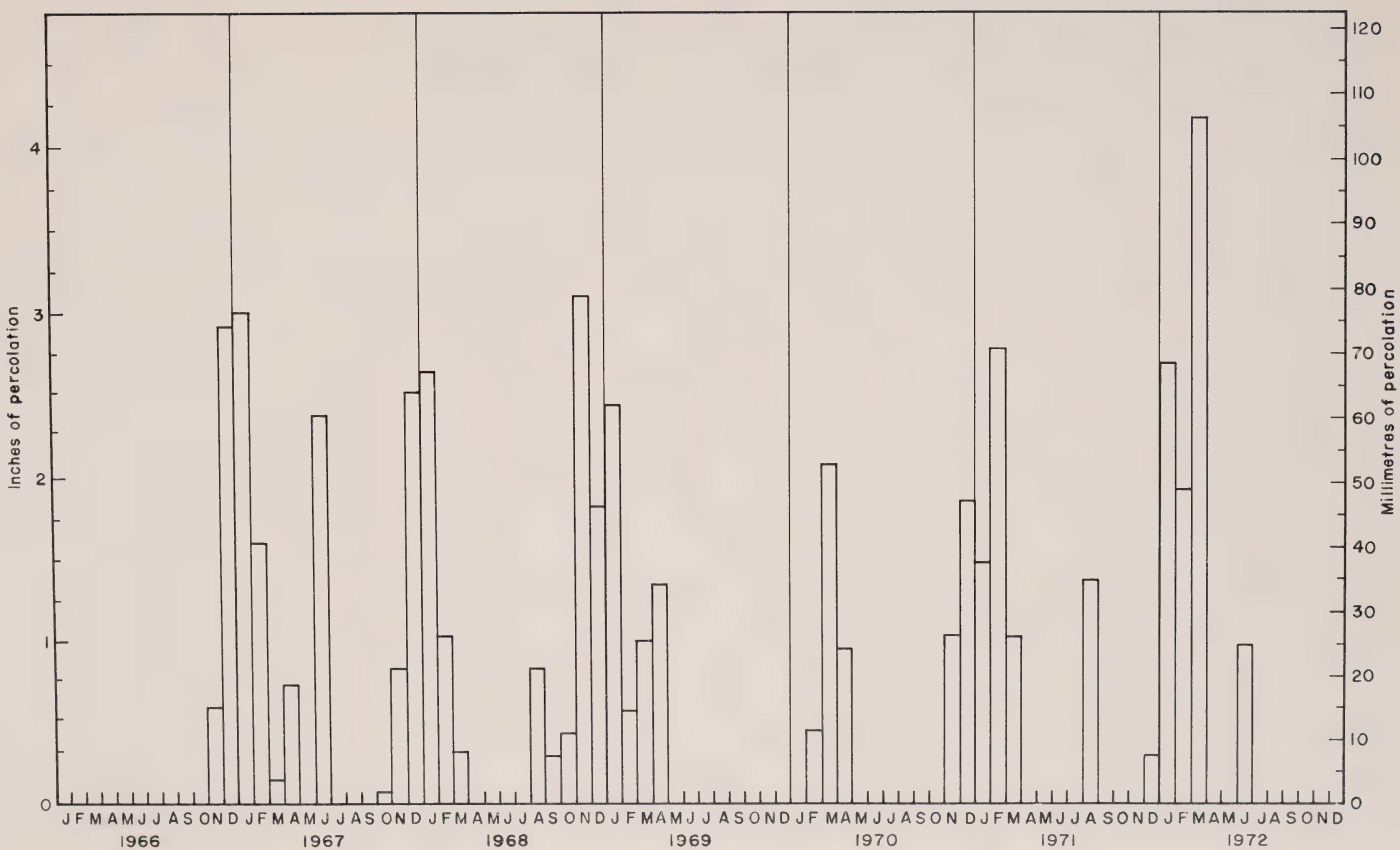


Figure 12. Estimates of the mean monthly percolation for the water years 1967 – 1972 in the Blue Springs Creek basin, using model.

The percolation estimates obtained by this analysis must be considered to be approximate values, due to the number of assumptions made using this method. In particular, the model did not represent conditions in the summer very well. The model predicted that the percolation in the summer was generally zero, and during the summer the mean monthly evapotranspiration was usually larger than the precipitation. However, a heavy rainfall during the summer would probably give some percolation. To model these short events would require reducing the time step to about a day.

The recharge can be estimated by using the ground water balance (Equation 9) , if the ground water storage can be determined. The change in ground water storage was approximated as:

$$\Delta GWS = \Delta \overline{WL}.SC.A \qquad (10)$$

where ΔGWS is the change in ground water storage,

$\Delta \overline{WL}$ is the change in mean water level in the area,

SC is the mean coefficient of storage,

and A is the basin area.

Combining equation 10 and 9:

$$R = B + \Delta(\overline{WL}).SC.A \qquad (11)$$

The mean water level in the basin was estimated by taking the mean level in the six observation wells representing discharge and recharge areas in the basin.

During some of the summer the recharge to the ground water is zero. During these periods the hydrographs of the observation wells can be seen to decline approximately exponentially, without any significant rise in water level caused by recharge. By inspection of the observation well records, 14 summer months (Figure 13) were identified when the recharge appeared to be zero.

If the recharge is zero, Equation (11) can be restated as:

$$SC = \frac{-B}{\Delta WL.A} \qquad (12)$$

Using Equation 12 for the 14 summer months when the recharge was apparently zero, the value of the coefficient of storage was found to vary between 0.03 to 0.05, with a mean of about 0.04. This mean value of storage coefficient was then used in Equation 11 to get the monthly recharge values for the years 1966 to 1972. These values have been plotted on Figure 13 and are listed in Appendix 1. For a few months the recharge was calculated to be negative; these values were set to zero, and the negative value was subtracted from the following month's recharge. For several of the summer months when the recharge was considered to be zero from inspection of the well hydrographs, a small recharge value was obtained by using Equation 11. These were due to the approximations

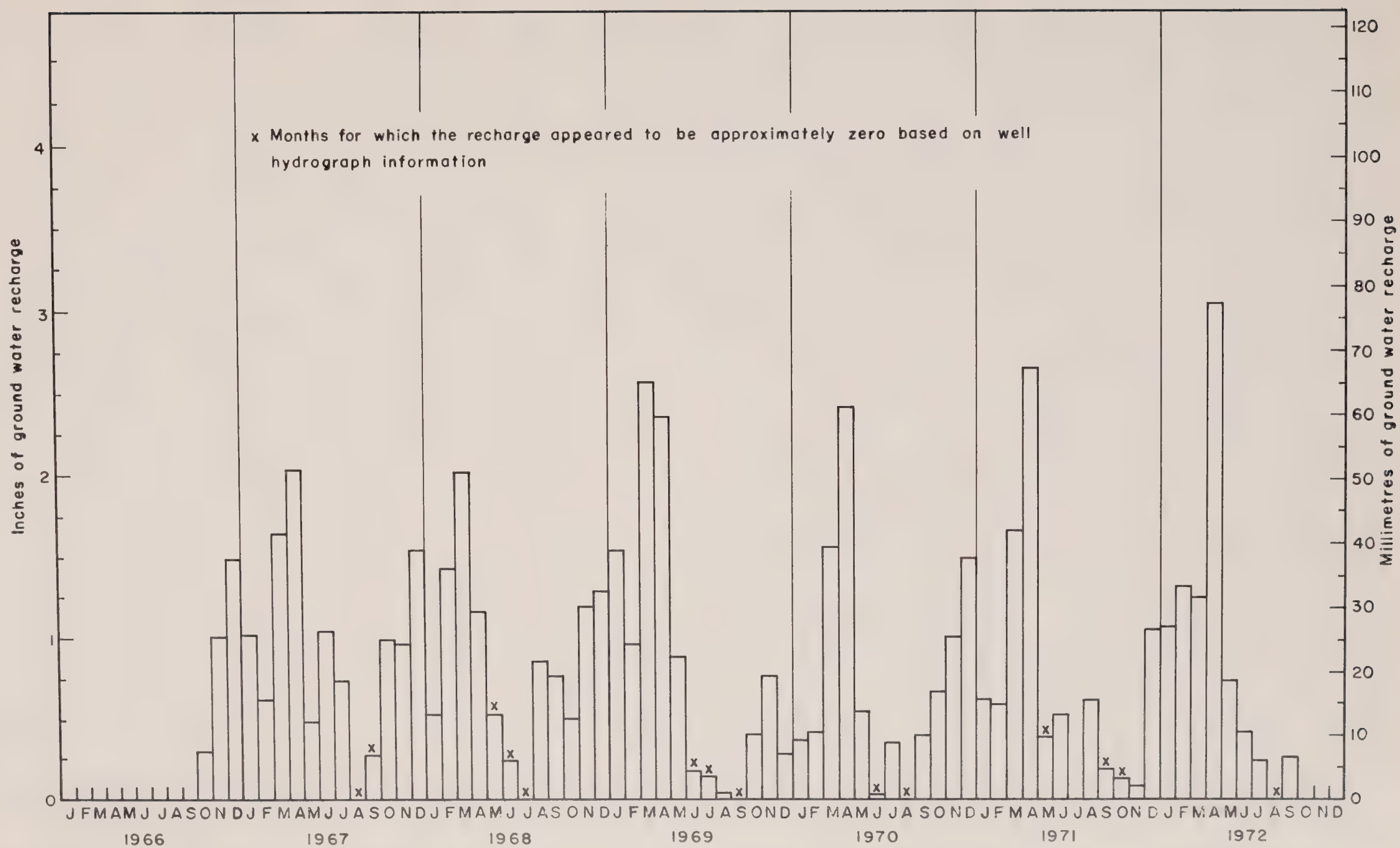


Figure 13. Estimates of the mean monthly ground water recharge for the water years 1967 – 1972 in the Blue Springs Creek basin, using model.

inherent in the method, and for consistency these figures were left at the value calculated by using Equation 11.

Comparing the percolation (Figure 12) with the recharge (Figure 13), it can be seen that the percolation is considerably more variable than the recharge. In addition, the percolation peaks are about one month ahead of the high recharge figures. The lower soil zone, therefore, acts to smooth and delay the percolation before it reaches the ground water.

Ground Water Discharge Zones

Ground water discharge zones were investigated by three different methods. Discharge and recharge zones along the stream were found by streamflow measurement. Other discharge zones were found by examining the data from the three piezometers and the general piezometric map. The methods and results are discussed in the following subsections.

Discrete recharge and discharge zones .... In temperate basins streams are usually gaining, but in karst basins the streams often recharge aquifers and receive ground water discharge at different points along their course. To estimate the amount of recharge and discharge in different reaches of Blue Springs Creek, the streamflow was measured at a number of points along the channel, on December 13, 1973; at this time flows were relatively low and receding. These measurements are shown in Figure 14 (in pocket), both in cfs and in inches of runoff per year. To compare the reaches of the stream that may be losing water to aquifers and those that gain ground water, a theoretical streamflow was calculated for each measuring point by assuming that the runoff was constant over the whole basin and was equivalent to 10.5 inches per year as measured at the downstream station. If runoff were evenly distributed all reaches would be supplying the theoretical streamflow. The differences between measured and theoretical values were used as an indicator of recharge and discharge areas and are shown in Figure 14.

The major area of losing flow appears to be the reach in the 4th Concession, Nassagaweya Twp., where there is a difference of approximately 4.3 cfs between the actual and theoretical streamflow. This water enters the ground water and it can be seen from Figure 14 that a comparable quantity of water discharges into Blue Springs Creek farther downstream. Other reaches of low runoff can be seen in the southerly part of the basin. These may be due to local infiltration to underlying aquifers or to ground water flowing to more distant reaches of the streams. All other reaches were found to be gaining, with runoff greater than the theoretical runoff.

Field surveys were carried out to check for sinks and resurgences along Blue Springs Creek and throughout the basin. A number of these sites were identified and are shown in Figure 20. Several streams sink in sand and gravel along the streambed, although no discrete sink points were seen that carried water all year.

There are a number of springs in the basin (Figure 20), most of which are small. One significant spring in the area can be found on the Boy Scouts Camp property, as the Blue Spring. This spring was discharging 0.47 cfs on December 13, 1973, which would correspond to a runoff from approximately 0.6 square miles, using the runoff rate at Eden Mills. The topographic basin area contributing to this spring is only about 0.05 square miles, and from the piezometric map (Map 7), it can be seen that the source of this spring probably includes the higher land to the southeast.

Recharge and discharge areas from well piezometer data .... The recharge and discharge areas were also investigated near the piezometer nests in the basin. A piezometer nest consists of a well containing a number of tubes which measure the hydraulic head in aquifers at different depths. From the static levels in each tube, the vertical component of ground water movement can be inferred. Three piezometer nests are located in the basin and are shown in Figure 15 (in pocket) as wells BS-1b, 4b, and 5 containing 4, 2 and 5 piezometer tubes respectively. The static water level measurements for the wells during 1972 are plotted in figures 16, 17 and 18. From these graphs, the average level in each piezometer tube in 1972 was calculated and converted to relative static levels, which are shown in Figure 19. The inferred vertical flow directions, based on the mean static level in the piezometers, are also shown in Figure 19.

In well 1b, located in the northern part of the basin, the predominant flow direction is upwards, indicating a discharge zone. The discharge of ground water causes a marsh to be formed in the vicinity of the well. This well, however, is near both the topographic and ground water divides (Figure 15) where, under normal conditions, the ground water would be expected to be flowing downwards. The lower aquifers near this well must be recharged from areas of higher elevation, probably from the high region to the north of the basin. Further information on the type of recharge can be obtained by inspection of the piezometer levels, Figure 16. During dry periods when the water levels are falling, the four piezometer tubes indicate that the water is flowing upwards and discharging at the surface. During wet periods, however, when the water levels are rising, the second deepest piezometer tube tapping a gravel bed at the 128-140 foot level, has the highest head. This bed must be connected to the recharge area by a relatively permeable path, while the aquifers at other levels must be more localized, and have a poorer connection to the recharge areas.

In well number 5, the upper piezometer tube had been filled with gravel to the 9.4 foot depth, and during much of the year no readings could be taken as the water level was below 9.4 feet.

However, when the static level is high enough for readings to be taken, the level is always at least three feet higher than the level in the other piezometers and this difference was assumed to exist throughout the year. From Figure 19, the flow direction is predominantly downwards, indicating a recharge zone.

Well number 4b, is also in a recharge area. The flow direction, as shown in Figure 19, is downwards.

Delineation of discharge zones on the basis of water level information .... In order to determine the approximate boundaries between the regional discharge and recharge zones from the bedrock aquifer, the following information was used.

A) The piezometric surface map (Map 7) was compared to the topographic elevations. The areas where the piezometric surface was at or above the ground surface indicate discharge areas;
B) The topographic map (Map 1) was inspected for marshy areas;
C) The soils map (Map 6) was inspected for marsh and Mesisol soil types. These areas occur either where ground water is being discharged or in depressions where the drainage is poor;

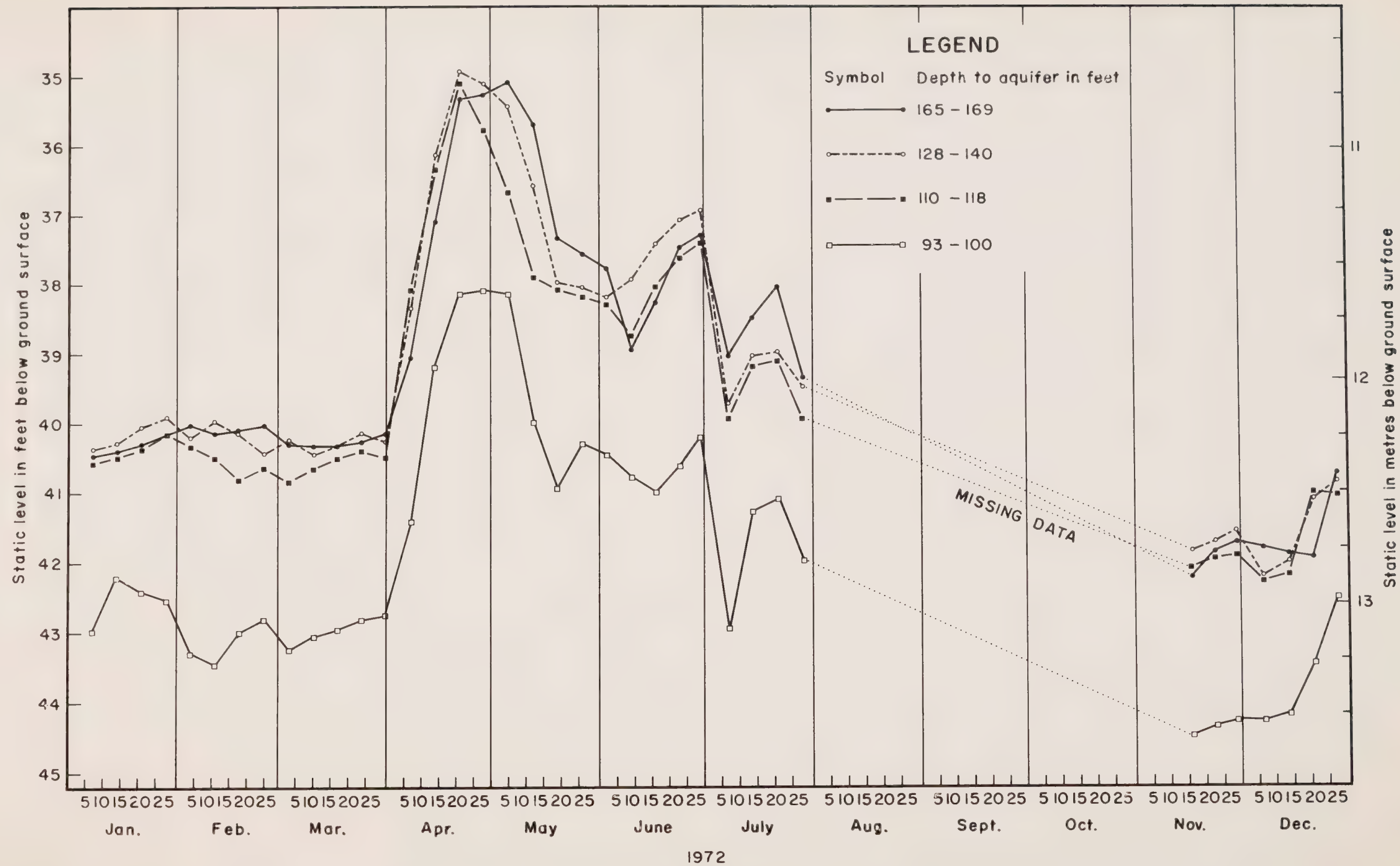


Figure 16. Static water level in well 1b during 1972 in the Blue Springs Creek basin.

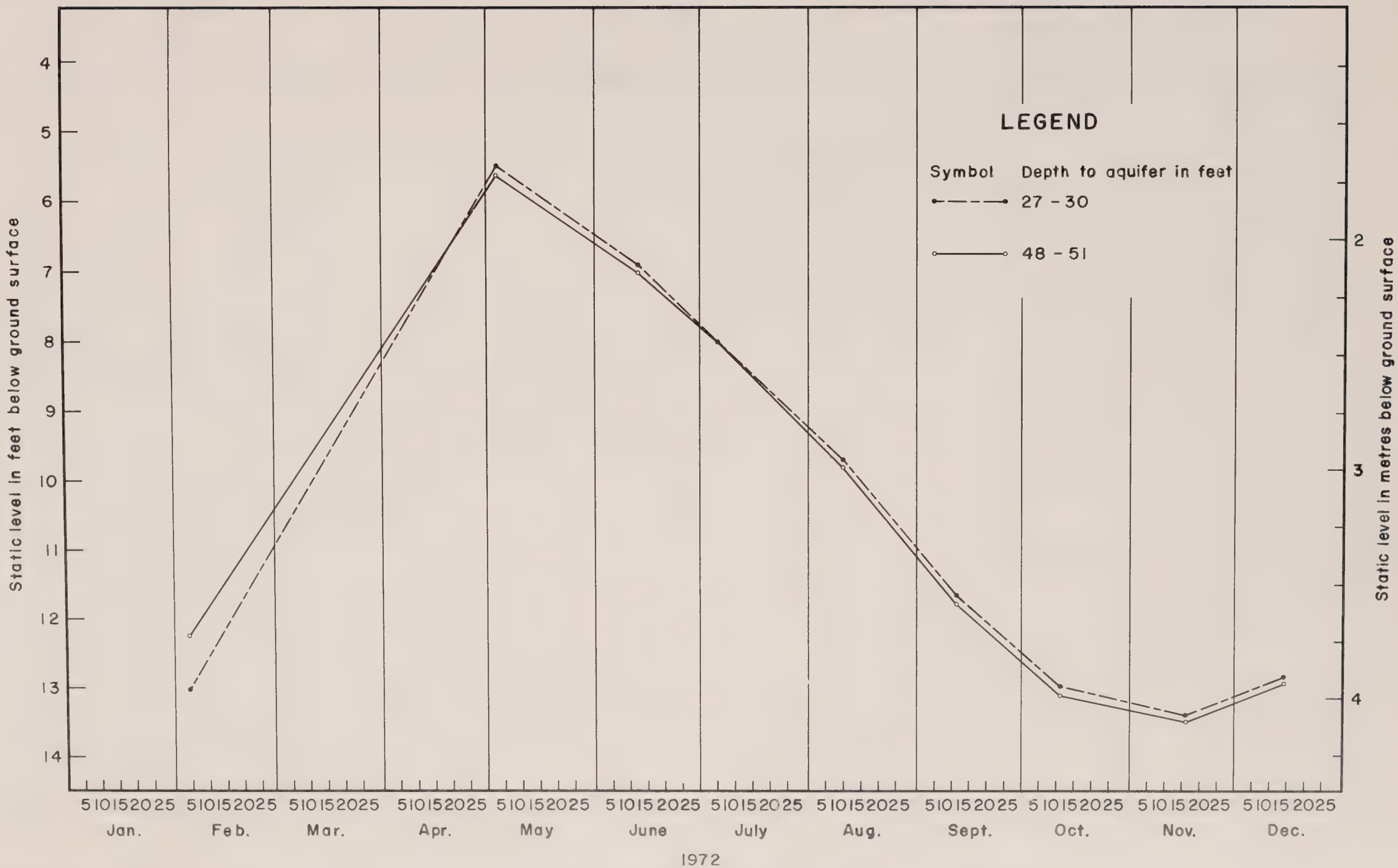


Figure 17. Static water level in well 4b during 1972 in the Blue Springs Creek basin.

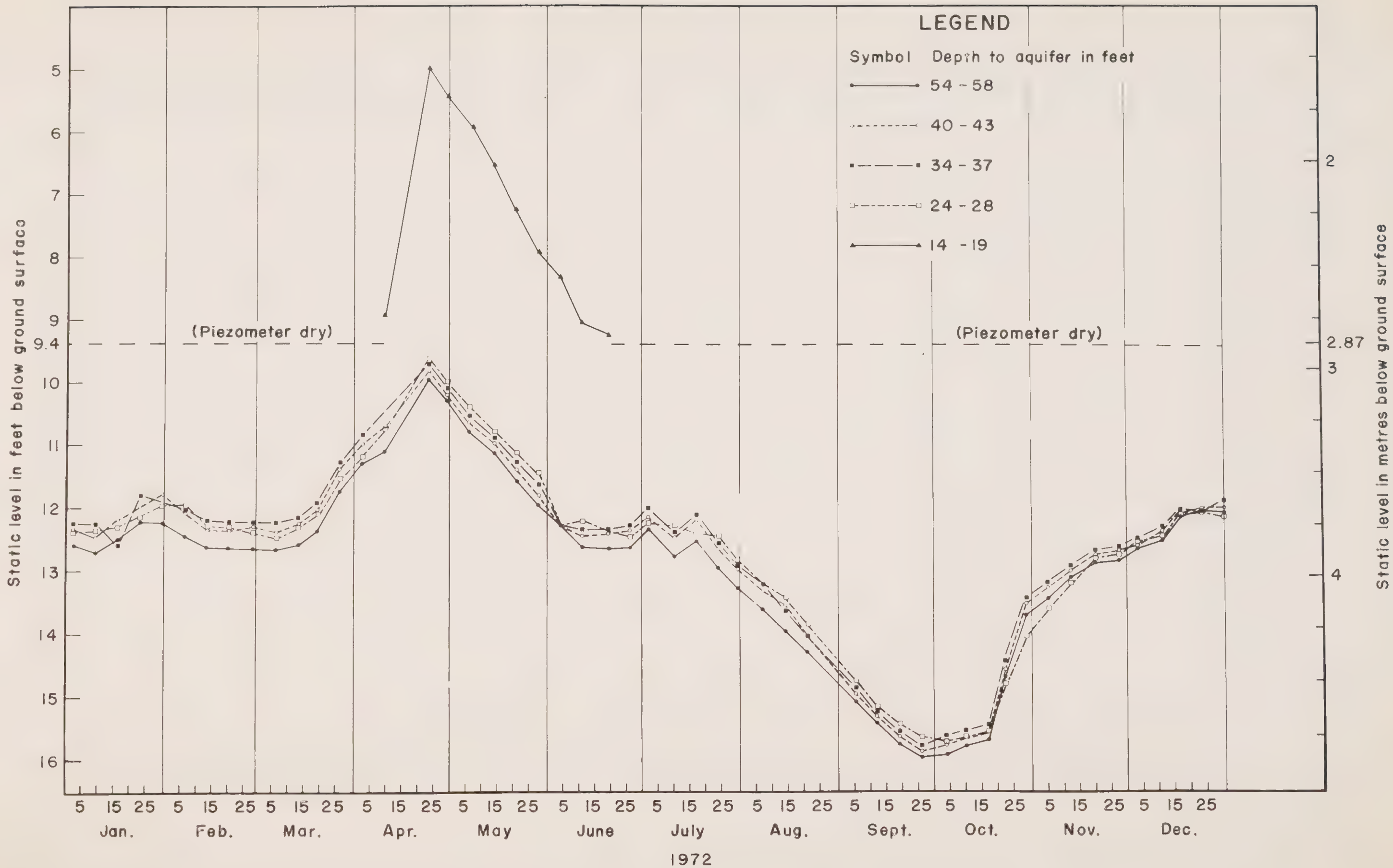


Figure 18. Static water level in well 5 during 1972 in the Blue Springs Creek basin.

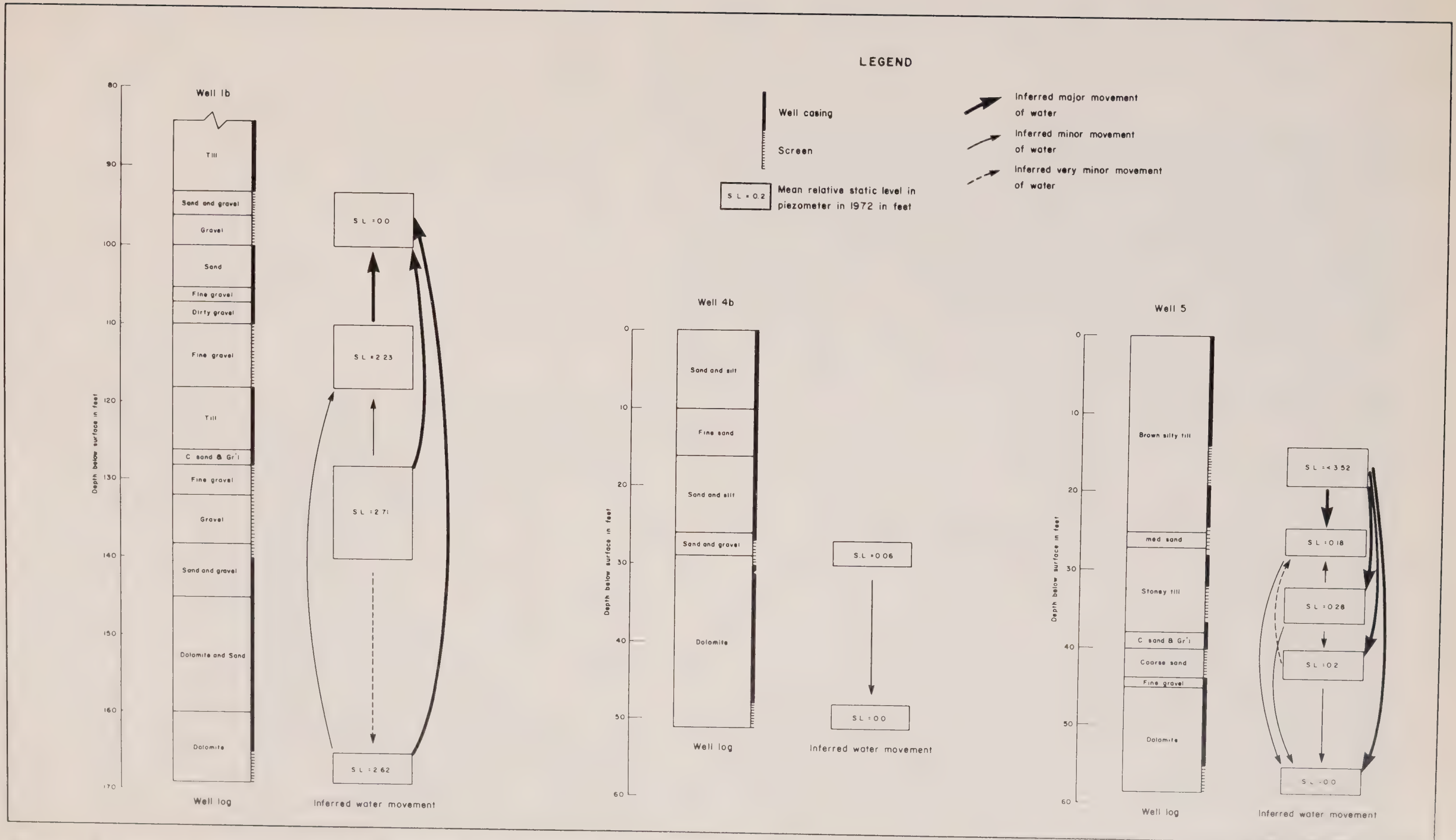


Figure 19. Well logs and inferred vertical flow directions in piezometer wells Ib, 4b and 5 in the Blue Springs Creek basin.

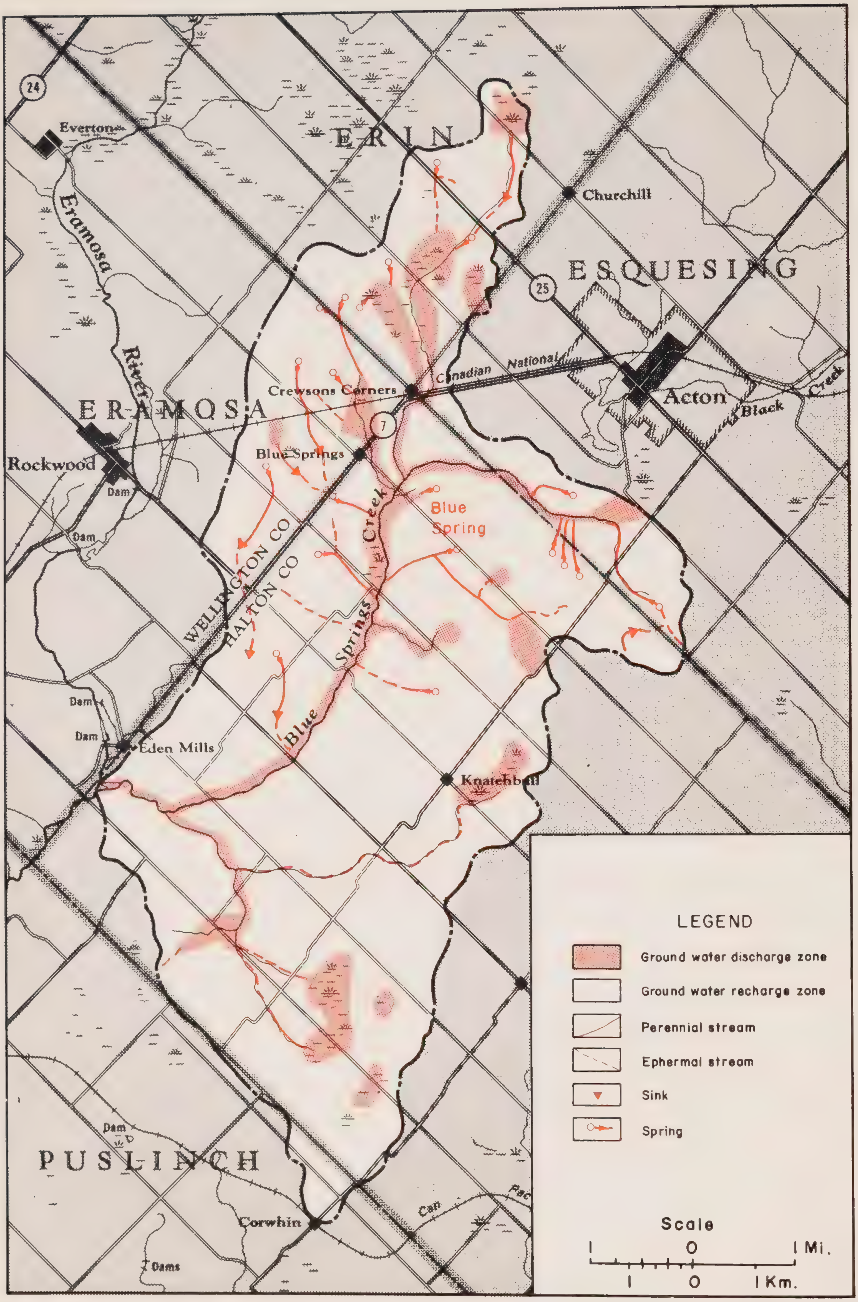


Figure 20. Approximate discharge and recharge zones in the Blue Springs Creek basin.

D) The air photographs of the region were studied for moist areas. These marshy areas were found to be darker on the aerial photographs and could be distinguished from the lighter, cultivated areas;

E) The piezometric well data (Section 3.3.2.2) were used;

F) In the field, springs were identified and located. Much of the field locating was carried out by Dickinson and Whiteley (1970);

G) The chemistry of the ground water was used to estimate recharge and discharge zones. This method is described in Section 5.5.

On the basis of the above evidence, the discharge and recharge zones were estimated and have been marked on Figure 20. The area of the discharge zones was measured at 1.9 sq. miles, (4.9 sq. km) or about six per cent of the basin area.

STREAMFLOW AND BASEFLOW CONDITIONS

The mean annual streamflows for the calendar years 1967 to 1971 are given in Table 7 for three stations in the Blue Springs Creek basin. The stations are the Cedar Grove and Boy Scouts Camp stations maintained by the University of Guelph, and the federal gauge 2GA-31. The station locations are shown in Figure 15. The topographic drainage areas above each station, the runoff per year, the annual precipitation and the percentage of runoff to precipitation are listed in Table 7.

The runoff from the federal gauge and the Boy Scouts Camp gauge is approximately 40 per cent of the precipitation. The runoff from the Cedar Grove gauge however, is only 23 per cent of the precipitation; this lower figure is due to the relatively smaller amount of ground water discharge in the Cedar Grove basin.

Ground Water Discharge Based on Hydrograph Separation

In any hydrograph, such as the one illustrated in Figure 21, it is generally considered that there are four major components of the streamflow (Linsley et al, 1949). These are:

a) Channel precipitation, which falls directly on the stream channel.

b) Surface runoff, which is conventionally regarded as the water that flows over the surface of the land. This is the major component of flood flows.

c) Interflow, which is the water that flows laterally through the soil without reaching the water table.

d) Baseflow, which is derived from the ground water reservoir. This is the major component of dry weather flows.

In practice, the separation of the hydrograph is difficult as there are no strict divisions between the components. Thus, some water may flow on the surface initially, then in the soil for a short distance, and back over the surface before reaching the stream network. This water may be classified either as interflow or surface runoff.

For this study, the streamflow was separated into the baseflow and the sum of the other three components, which together were called the overland flow. This analysis was carried out by conventional baseflow separation techniques (Gray and Wigham, 1970).

The hydrograph of streamflow, during periods when all discharge is

TABLE 7. MEAN ANNUAL STREAMFLOWS IN THE BLUE SPRINGS CREEK BASIN, 1967-1971

		Cedar Grove			Boy Scouts Camp			Federal Gauge		
Basin Area Sq. Miles (Sq. km)			2.05 (5.31)			8.81 (22.81)			18.12 (46.90)	
		Mean Annual Streamflow								
Calendar Year	Precipitation Inches/Yr (mm/yr)	(cfs)	Inches/Yr (mm/yr)	(% of Precip.)	(cfs)	Inches/yr (mm/yr)	(% of Precip.)	(cfs)	Inches/yr (mm/yr)	(% of Precip).
1967	36.45 (926)	1.22	8.1 (206)	22.2	9.3	14.5 (368)	39.8	19.9	15.0 (38)	41.3
1968	38.20 (970)	1.46	9.8 (249)	25.6	10.0	15.5 (394)	40.6	21.1	15.9 (404)	41.7
1969	30.50 (775)	1.44	9.6 (244)	31.4	8.6	13.4 (340)	43.9	20.4	15.4 (391)	50.5
1970	33.34 (847)	0.78	5.2 (132)	15.6	6.5	10.1 (256)	30.3	15.3	11.5 (292)	34.5
1971	34.27 (871)	1.09	7.3 (185)	21.3	8.14	12.7 (323)	37.1	19.0	14.3 (363)	41.7
Mean	34.55 (878)	1.20	8.0 (203)	23.2	8.51	13.2 (335)	38.3	19.14	14.4 (366)	41.7

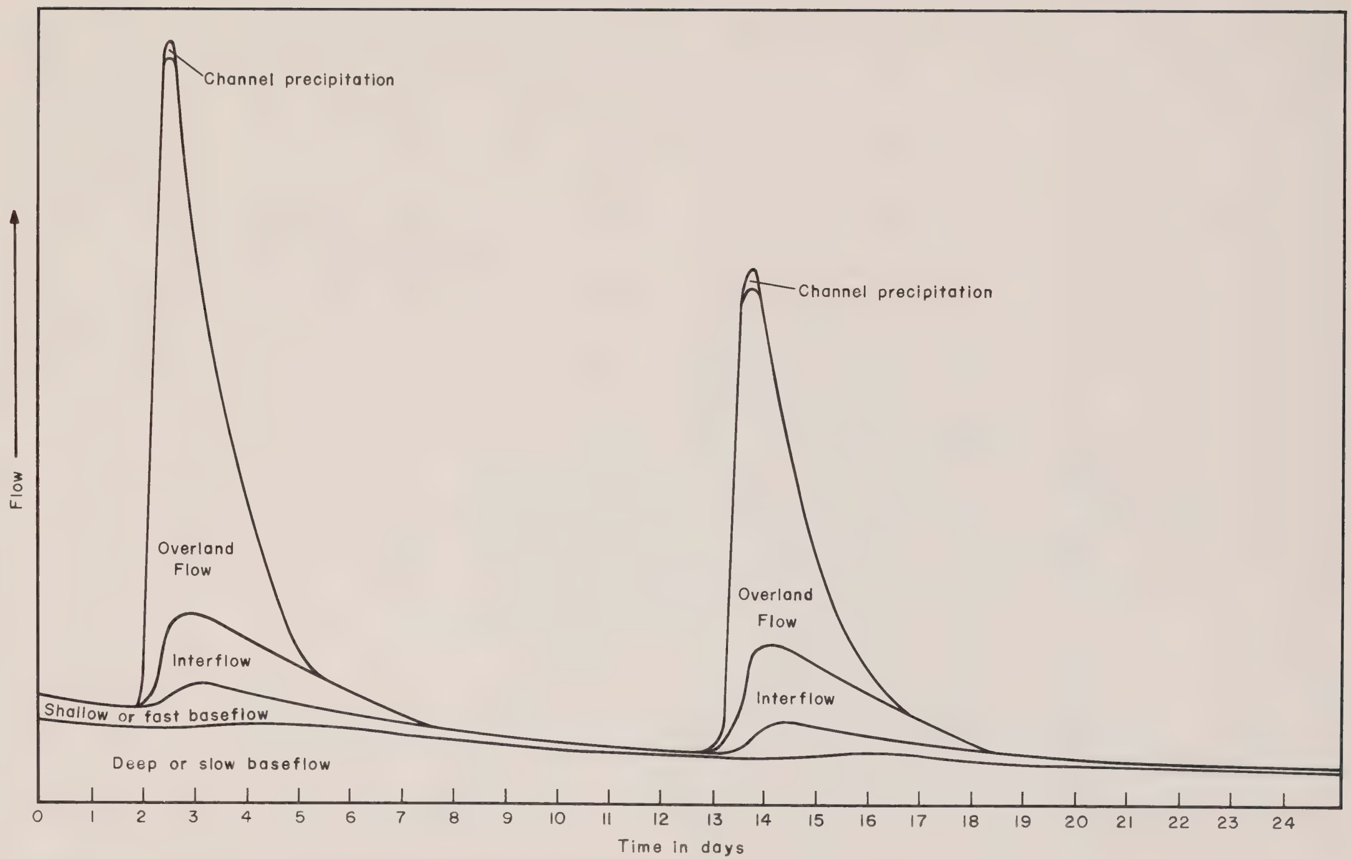


Figure 21. Conceptual hydrograph showing the components of streamflow.

derived from ground water sources, is known as the baseflow recession curve. Several theoretical equations have been proposed to describe the recession curve. They are generally of the form (Barnes, 1940):

$$Q_t = Q_o K_r^{-t} \qquad (13)$$

where:

Q_t = discharge at any instant,

Q_o = discharge at some initial time,

K_r = recession constant,

and t = time interval between Q_o and Q_t.

A baseflow recession curve within the period extending from February 15 to March 12, in the year 1969, at the federal gauging station, was selected and assumed to represent the ground water outflow in the basin, as shown in Figure 24. A winter recession curve was selected because evapotranspiration losses to the atmosphere during that period were assumed negligible. Substituting the flow values from the recession curve from Figure 24 into Equation 13 gives a recession constant K_r of 0.989, or a 'half life' for the recession of 64 days. Thus the baseflow after a 64-day period will be reduced by one half if the aquifer has not been recharged.

The slope of the recession curve was drawn on the stream hydrographs (Figures 22 to 24) showing total runoff on semi-logarithmic paper. The recession curve plotted in this manner appeared to fit the falling stages of the hydrographs for the time periods 1966-1969. In order to determine the limits of the baseflow recession curves on the stream hydrographs, observation well (2b) was used as an index well which was assumed to reflect the natural dewatering and recharge effects on the ground water reservoir.

Figures 22 to 24 illustrate the manner in which the stream hydrograph was separated into two components, surface runoff and ground water discharge. The daily ground water discharges beneath the recession curves were then added to give the total monthly ground water discharge; Table 8 summarizes the monthly ground water discharge values.

The lowest and highest monthly ground water discharge, based on the three year period, were 8.5 and 25.6 cfs, respectively. The mean ground water discharge was 14.0 cfs, which was 68% of the total streamflow.

Specific Yields from Different Sub-basins in the Blue Springs Creek Basin

When an aquifer is drained under the action of gravity, the amount of water yielded is less than the total volume of the aquifer material drained (Wisler and Brater, 1959). The quantity of water that is yielded from the aquifer is called the specific yield, and is usually expressed as a percentage of the total aquifer volume. During dry periods when the aquifer is not being recharged, and if evapotranspiration and pumpage from the aquifer are small, the specific yield can be given as:

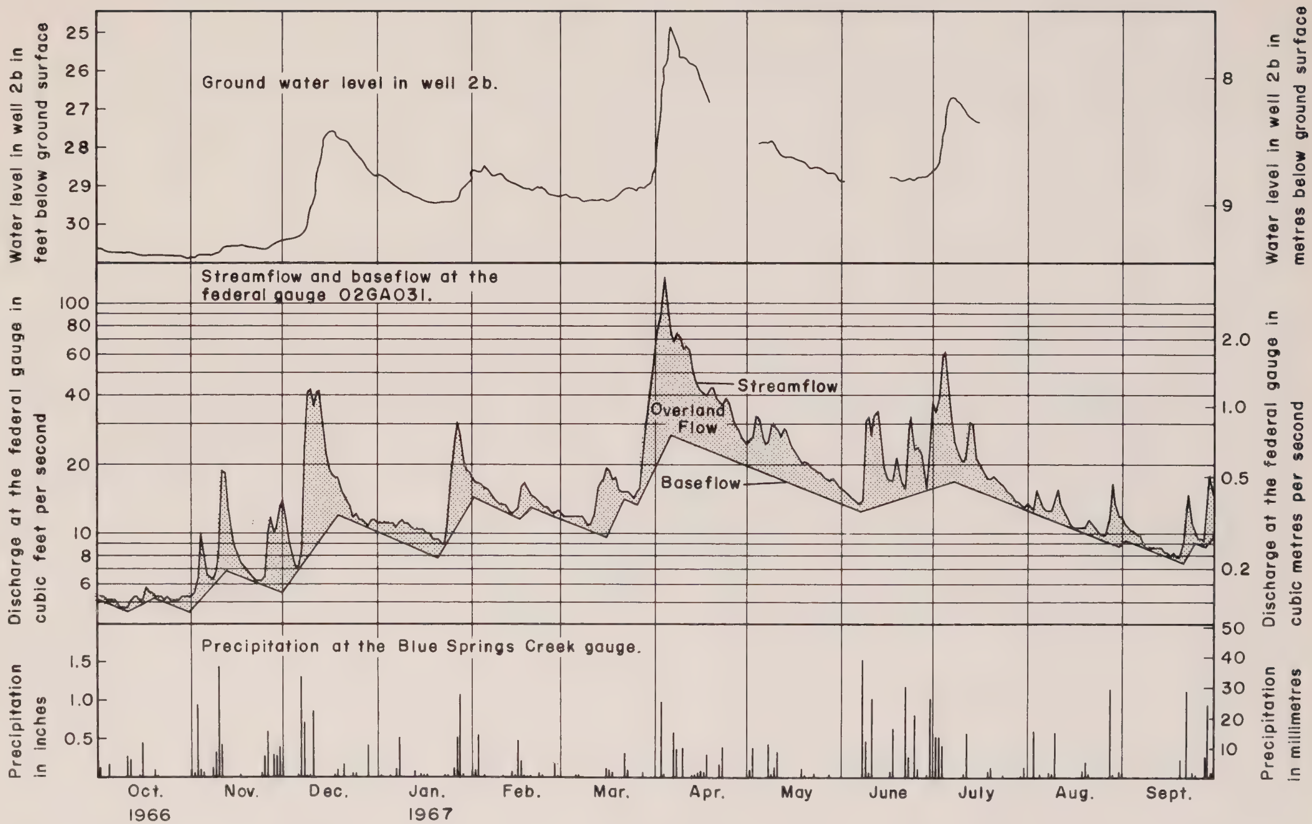


Figure 22. Daily streamflow and baseflow at the federal gauge O2GA031, ground water level at index well 2b and precipitation in the Blue Springs Creek basin for the water year 1966-67.

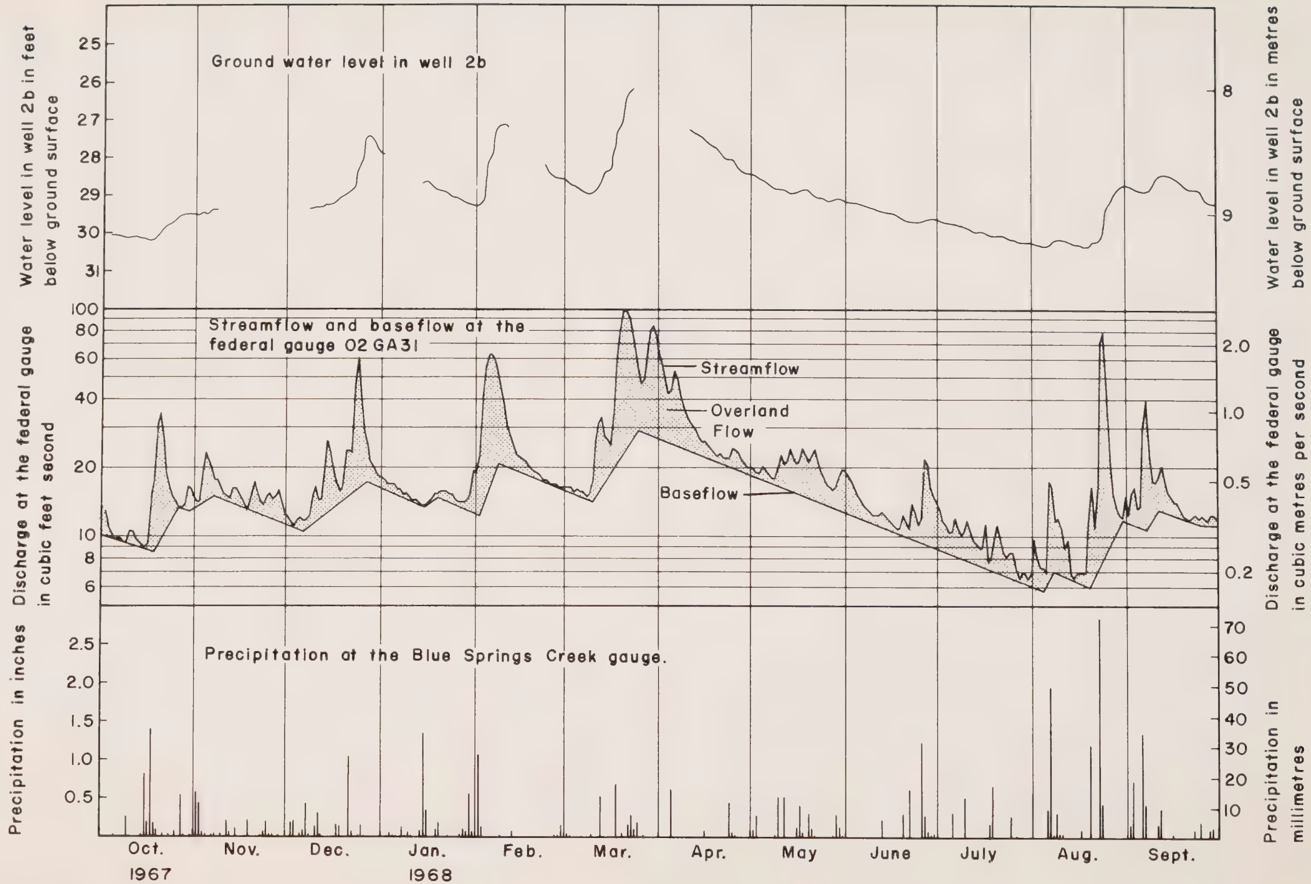


Figure 23. Daily streamflow and baseflow at the federal gauge 02GA031, ground water level at index well 2b and precipitation in the Blue Springs Creek basin for the water year 1967 - 68.

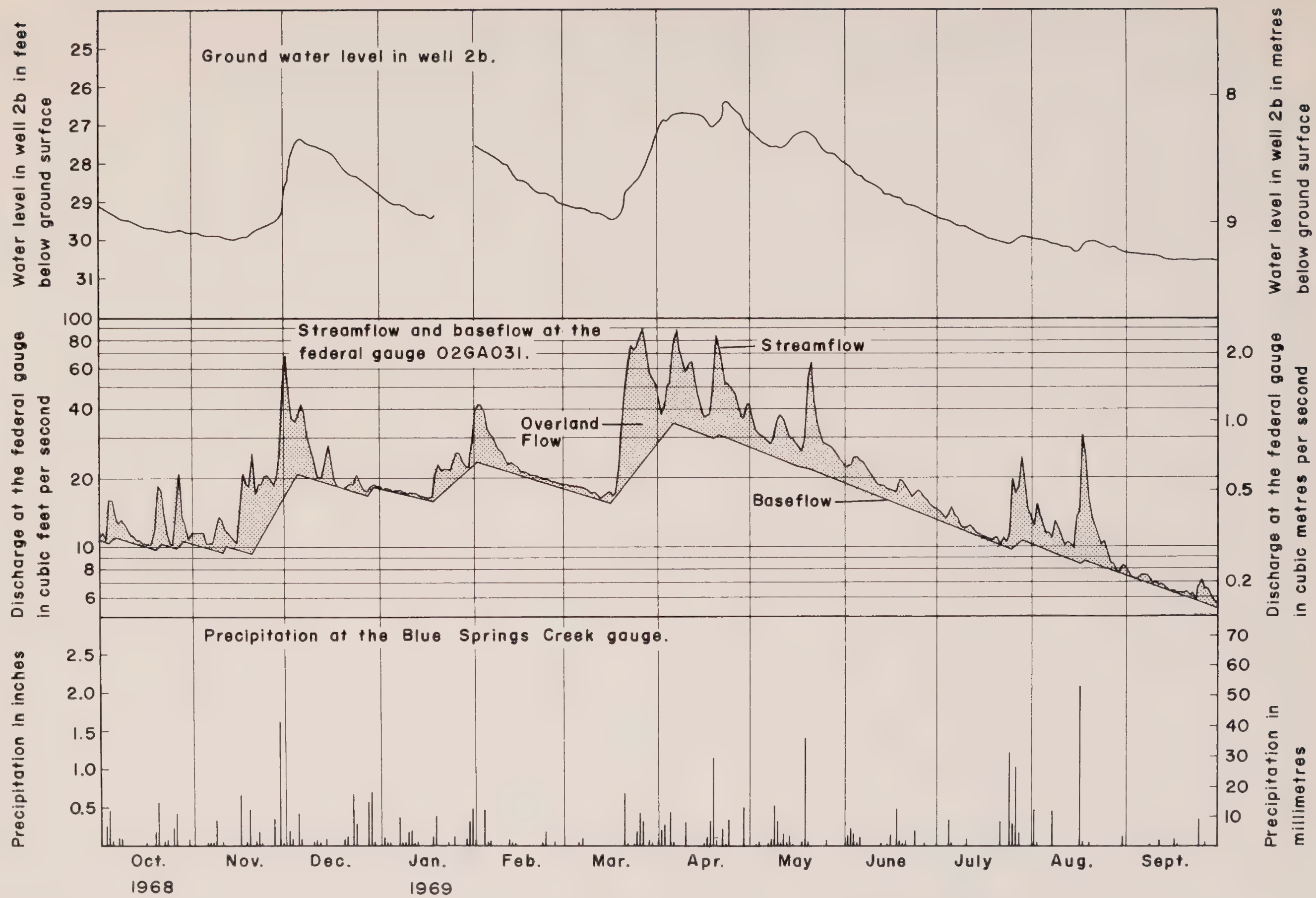


Figure 24. Daily streamflow and baseflow at the federal gauge O2GAO31, ground water level at index well 2b and precipitation in the Blue Springs Creek basin for the water year 1968-69.

TABLE 8. MONTHLY AND MEAN MONTHLY GROUND WATER DISCHARGE (IN CFS-DAYS) AT THE FEDERAL GAUGE 02GA031 IN THE BLUE SPRINGS CREEK BASIN

	Water Years			
Month	1966-1967	1967-68	1968-69	Mean
October	150.3	325.4	305.7	260.4
November	179.7	401.3	295.5	292.1
December	315.1	427.7	570.0	437.6
January	300.7	420.5	569.3	430.1
February	366.6	518.5	596.7	493.9
March	383.8	695.2	586.8	555.2
April	693.2	711.4	927.3	777.2
May	508.2	507.3	713.2	576.2
June	416.9	338.5	480.1	411.8
July	466.2	242.9	349.6	352.9
August	323.8	235.5	270.3	276.5
September	251.4	360.7	185.4	265.8
TOTAL	4355.7	5185.0	5849.9	5129.8
MEAN MONTHLY	362.9	432.0	487.4	427.2
Mean Baseflow cfs	11.9	14.2	16.0	14.0
Mean Stream-flow at 02GA31 cfs	18.5	20.8	22.6	20.6
Percentage of Baseflow to Streamflow	64%	68%	71%	68%

$$y = - \frac{Q\Delta t}{\Delta hA} \times 100 \qquad (14)$$

where y is the specific yield in per cent,

Q is the mean baseflow discharge during the period Δt,

Δh is the mean change in piezometric head over the area A.

The specific yield can be calculated for aquifers in a sub-basin if the mean decline in head and mean baseflow can be estimated.

The specific yield was estimated for the six sub-basins marked on Figure 15. These are the Cedar Grove, Boy Scouts Camp, federal gauge and the Blue Springs Creek basins above each gauge, and the Knatchbull and the Eden Mills sub-basins between the lower two gauges.

The calculations were carried out for the period February 15 to March 12, 1969 when the precipitation and evapotranspiration were assumed to be negligible so that the entire streamflow was taken to be baseflow. The change in head in the aquifer was estimated from the recording observation wells 1a, 2a, 2b, 3 and 4a and at the piezometer nest 5 (see locations on Figure 15). The heads in these six wells are shown in Figure 25. The mean change in head in any sub-basin was calculated by taking the average decline in head of the wells in or near that sub-basin. The wells considered for each sub-basin and the average decline in head are shown in Table 9.

The streamflows at the four gauging stations are plotted on Figure 26. The mean baseflows were calculated by averaging the daily flows. The baseflows in the Knatchbull and Eden Mills sub-basins were taken as the differences between the baseflow at the gauge downstream and the gauge upstream of the sub-basin. The specific yields were calculated by using Equation 14 for the six sub-basins, and are recorded in Table 9.

The specific yields in the sub-basins range from 7.2 to 12.7 per cent (Table 9). The lowest specific yields are from the Boy Scouts Camp and Cedar Grove sub-basins, which are in the northern portion of the area. The major portion of the ground water in these sub-basins is derived from the sandy till reservoir. The highest specific yields are found in the Knatchbull and Eden Mills sub-basins, which derive much of their ground water from the dolomite aquifer. However, as much of the dolomite aquifer is overlain by saturated till, the specific yields found in these sub-basins also represent the conditions in the overlying till. The sandy till aquifer is composed of kame, esker, outwash and alluvial sands and gravels and sandy till. The sands and gravels generally have a higher permeability than tills and could be expected to have a higher specific yield than the finer till (Todd, 1959), if conditions of thickness and degree of saturation were similar.

The proportion of the surficial sands and gravels (shown on Map 2) in each sub-basin is given in Table 9. With the exception of the Cedar Grove and Eden Mills sub-basins, the specific yield generally increases as the proportion of sands and gravels in the basin increases. In the Cedar Grove sub-basin, which has a small specific yield, the proportion of surficial sands and gravels is high, but these sands and gravels tend to be thin and are not very persistent. The shallowness of these deposits can be seen in the geological cross sections A-A' and B-B' of Fig. 5 (in pocket). On the other hand the Eden Mills sub-basin, which has a high specific yield, contains a small proportion of sands and gravels but

TABLE 9. SPECIFIC YIELDS FOR DIFFERENT SUB-BASINS IN THE BLUE SPRINGS CREEK BASIN, DETERMINED FOR THE PERIOD FEBRUARY 15 TO MARCH 12, 1969

Sub-basin	Mean Baseflow (cfs)	Sub-basin Area (sq. miles)	Observation wells in or Near Sub-basin	Mean Change in Head (feet)	Specific Yield (%)	Proportion of Sub-basin Area with Surficial Kame, Esker and Outwash Sand & Gravels (%)
Cedar Grove	1.25	2.05	1a	-0.57	8.2	34
Boy Scouts Camp	7.55	8.81	1a, 2a, 2b,	-0.92	7.2	18
Federal Gauge	18.5	18.12	1a, 2a, 2b, 4a, 5	-0.83	9.5	29
Blue Springs Creek	30.2	30.31	1a, 2a, 2b, 3, 4a, 5	-0.76	10.2	24
Knatchball	11.05	9.31	2b, 41, 5	-0.76	12.2	39
Eden Mills	11.7	12.19	3, 4a, 5	-0.59	12.7	18

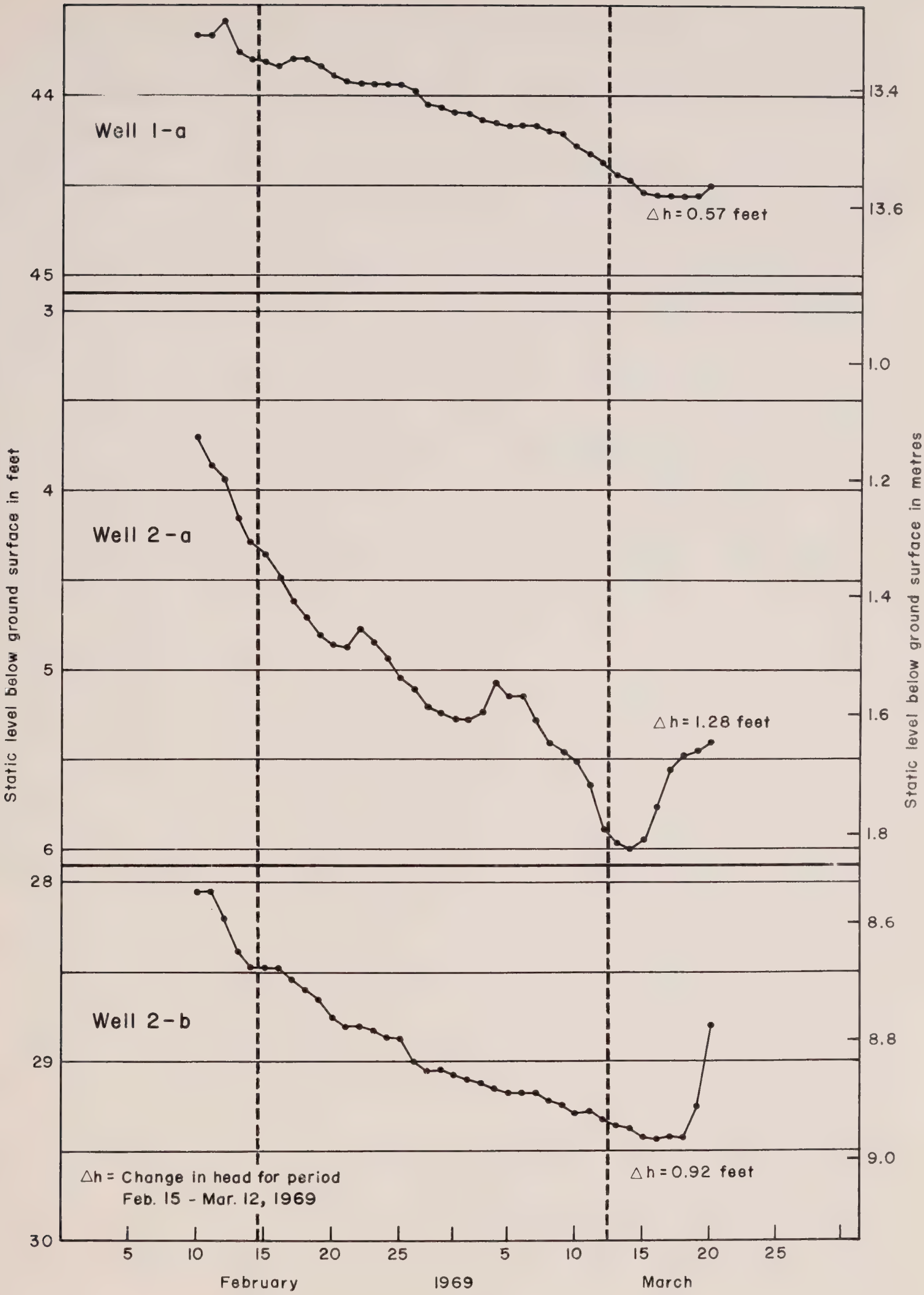


Figure 25a. Storage conditions in the sandy till reservoir as shown by observation wells 1a, 2a, 2b in the Blue Springs Creek basin, February 15 to March 12, 1969.

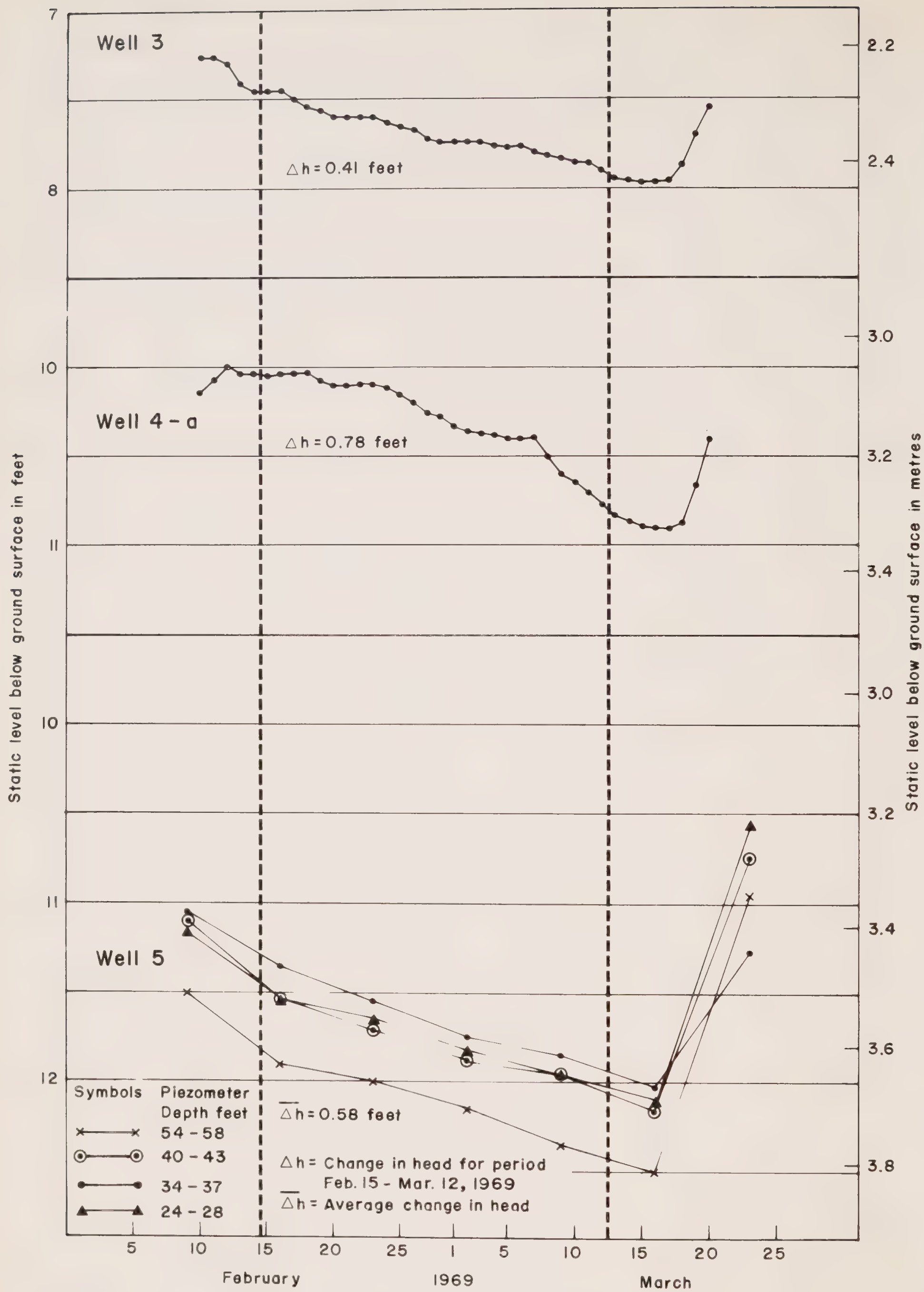


Figure 25b. Storage conditions in the sandy till reservoir as shown by observation wells 3, 4a and at the piezometer nest 5 in the Blue Springs Creek basin, February 15 to March 12, 1969.

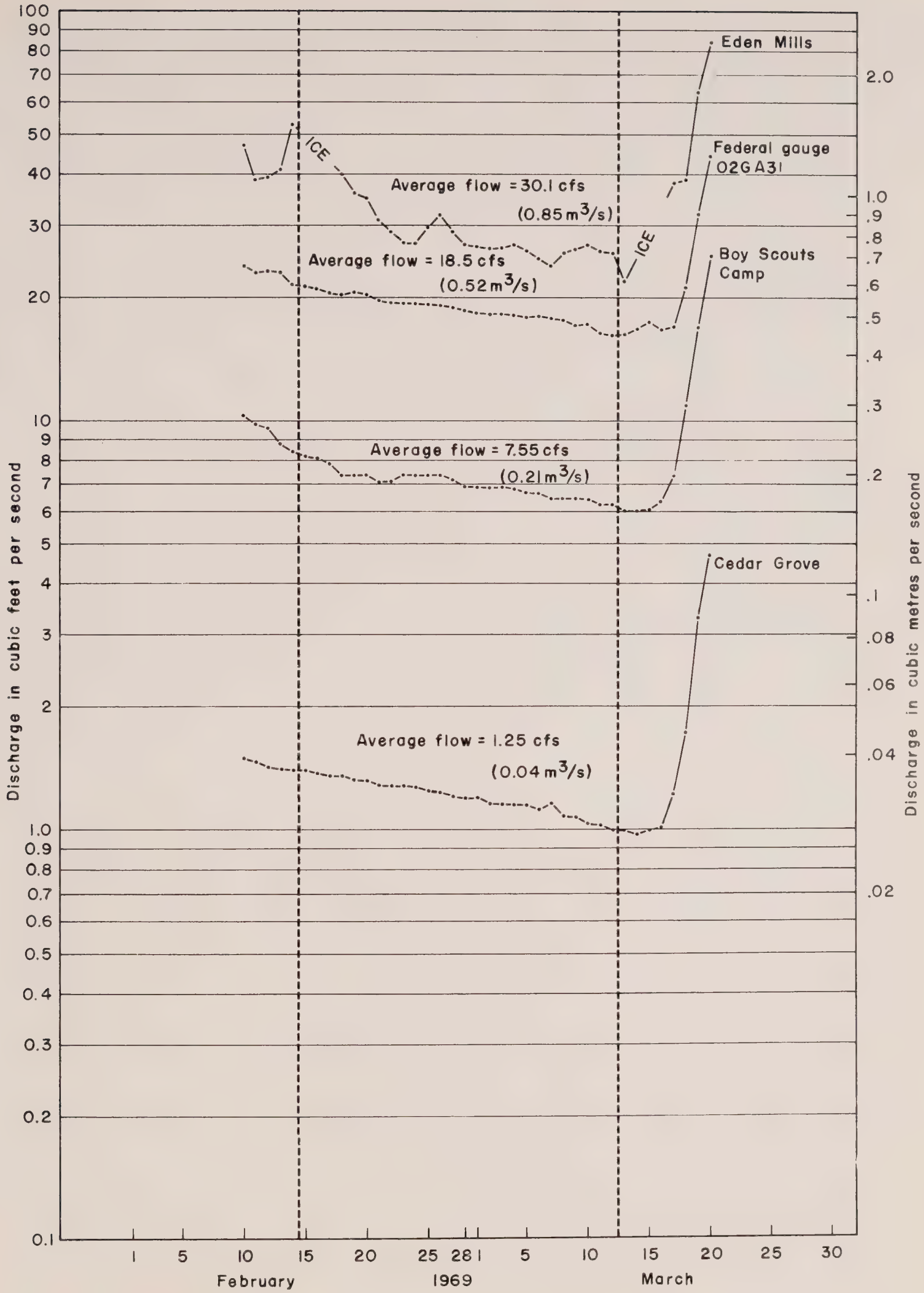


Figure 26. Daily discharges at the Cedar Grove, Boy Scouts Camp, federal and Eden Mills gauges for the period February 15 to March 12, 1969.

deposits are thick and persistent, as can be seen by cross-section G-G'. Therefore, it appears that large specific yields in a basin are related to large, thick and persistent areas of kame, esker, outwash and alluvial sands and gravels.

GROUND WATER MODELLING

As described in previous chapters, the transmissivities, coefficients of storage and recharge of an aquifer are important in understanding its flow pattern and defining its potential yield. In this chapter, two phases of ground water modelling are applied to evaluate and extend the results obtained by conventional techniques, and to provide a better understanding of the regional distribution of these aquifer characteristics and their interaction with the streams in the basin.

In the previous chapter, estimations of the transmissivity values were obtained, based on well record analysis and on ground water flow rates. Although these techniques yielded fairly consistent estimates of the transmissivity, they could not be used to obtain satisfactory figures for the regional transmissivity throughout the basin. Well record analysis provided values for the local conditions near the wells. Analysis of the ground water flow rate did provide regional transmissivity values; however, these were limited to only parts of the basin, due to lack of extensive data. One method of further deriving the regional variations of transmissivity is through the use of modelling. In the Blue Springs Creek basin, a steady state, finite difference model was used to estimate the regional transmissivity values; the figures obtained were then compared to those found in Chapter 3.

The coefficients of storage and recharge to the ground water aquifer are other important parameters in understanding the ground water regime. Both parameters, however, are difficult to determine by conventional techniques. A dynamic finite difference model was used to estimate the coefficient of storage throughout the basin and to verify the time-varying recharge rates to the aquifer calculated by water balance considerations.

One major concern in applying ground water models to simulate limestone or dolomite aquifers is the possible presence of karst development and the need to incorporate the associated flow characteristics into a modelling routine. As illustrated previously, if the karst conditions in a basin are well developed, solutional channels may transmit large quantities of water through the aquifer. If such an area is to be modelled accurately, the individual solutional channels have to be identified and located in the field in order for the simulation model to adequately account for their effect. If these channels, however, were not identified, the ground water model would not adequately simulate the aquifer characteristics. In the case of the Blue Springs Creek basin, the karst development has been shown to be of limited extent, the effects of the minor solution channelling, sinkholes and springs having no detectable effect on the piezometric water level surface of the area (Section 2.1.4). For the Blue Springs Creek basin, the ground water was modelled by assuming that the flows obeyed the Darcian flow equations. This appears valid, considering that the transmission of flow through solutional channels is minimal and only local. The model then was able to simulate the average aquifer characteristics on a distributed area basis, showing their general distribution and variation throughout the basin and its surrounding area.

DESCRIPTION OF THE MODELS

The programs used to solve the ground water flow equations by the

finite difference method are described by Smith, 1965, and Prickett and Lonnquist, 1971. A brief summary of the mathematics and the programs is presented here.

The generalized two dimensional differential equation of flow in an aquifer is given by:

$$\frac{d}{dy}\left(Ty\frac{dh}{dy}\right) + \frac{d}{dx}\left(Tx\frac{dh}{dx}\right) + I = S\frac{dh}{dt} \qquad (15)$$

where T_x, T_y are the transmissivity components in the x and y directions respectively,

I is the sink or source term, representing pumpage or infiltration,

h is the piezometric head in the aquifer,

S is the storage coefficient,

and t is time.

In the steady state case, when the head does not vary with time, the right hand side of the equation will be zero.

Equation 15 is the basic equation for all two dimensional ground water flow problems. There is no general solution for this equation, but approximate solutions can be obtained by a number of methods. In the finite difference method of solution, the differential terms in Equation 15 are replaced by their finite difference approximations. The aquifer being modelled is then discretized into a nodal network and the finite difference equations are solved at each node in turn.

In order to solve the problem more expediently, some physical idealizations have to be made (Figure 27). The aquifer is assumed to be thin and confined below by an aquitard, so that the flow in the aquifer is assumed to be horizontal. Infiltration can occur to the aquifer through the leaky confining layer above it and pumpage or spring discharge can occur at any node in the model. Recharge or discharge can occur through the permeable beds of the rivers which are assumed to have a constant head at all times. In practice, these assumptions appear not unduly restrictive.

The first step in the modelling process is to superimpose a grid network over the area to be modelled. The grid used for the Blue Springs Creek basin and the surrounding area is shown in Figure 28; it was set up unequally spaced with more nodes in the basin area where detailed head information is available.

At each node, all sources of water that can enter or leave the node are considered, as shown diagrammatically in Figure 29. There are seven routes that the water can enter or leave any central node (marked 'O' on Figure 29). These are: to or from the four adjacent nodes, to or from the river, to or from storage and by direct pumpage or recharge. By altering the theoretical head at the central node, it is possible to balance the amount of water entering this central node with the amount leaving. By repeating this process for every node in the basin, on an iterative basis, the theoretical head distribution throughout the aquifer can be calculated.

In the steady state case, the model uses the "overrelaxation"

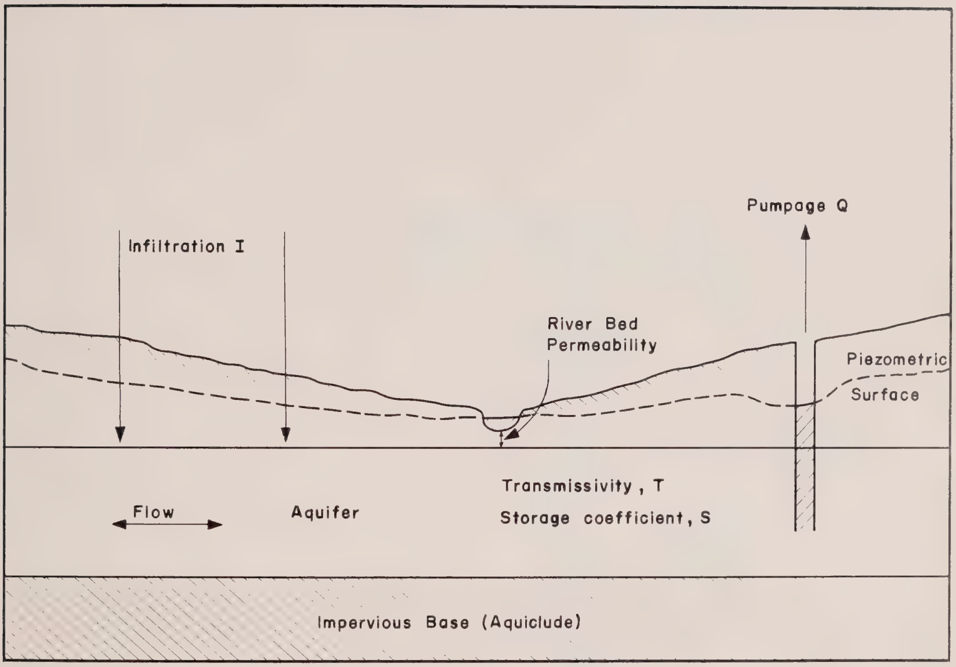


Figure 27. The idealized aquifer system modelled.

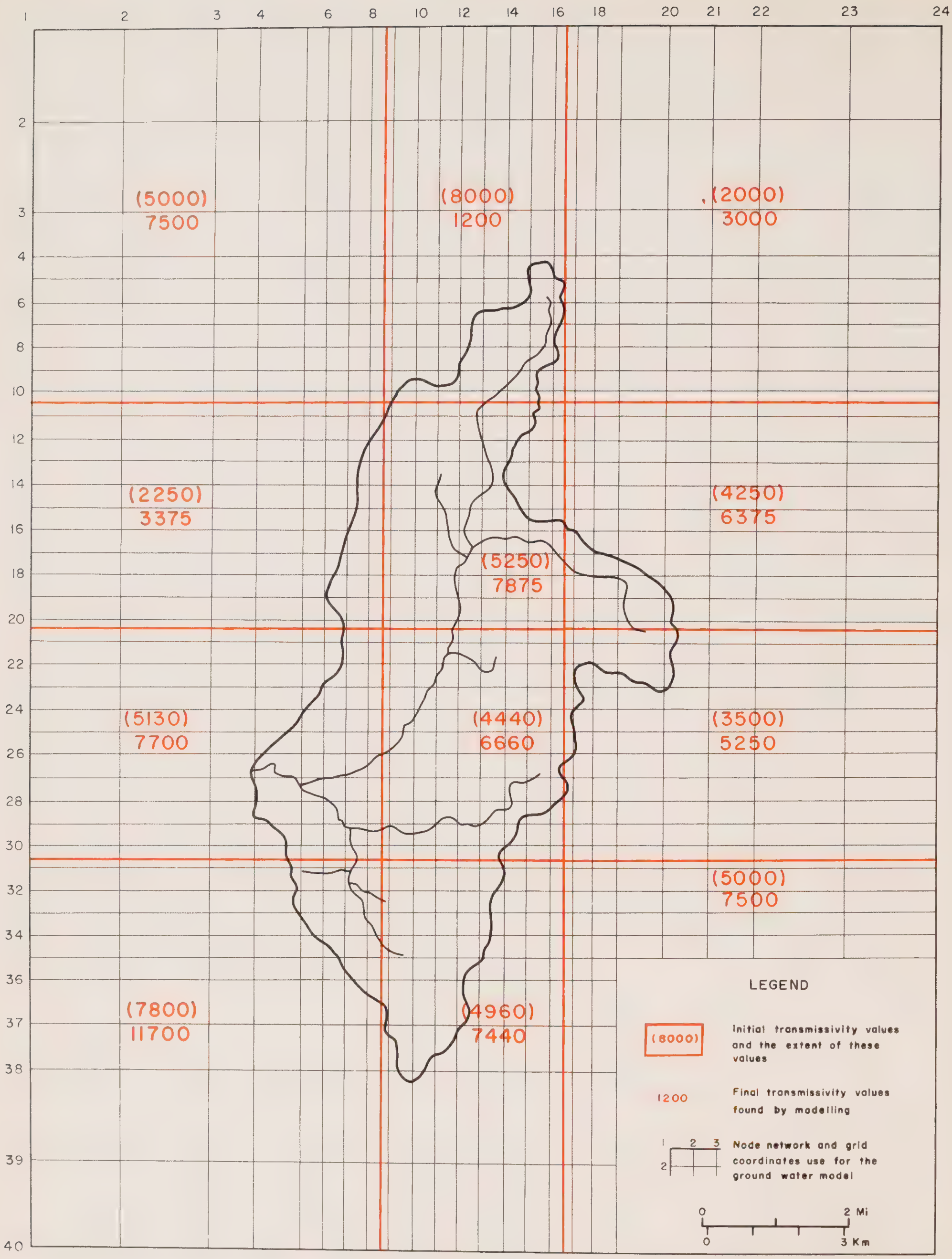
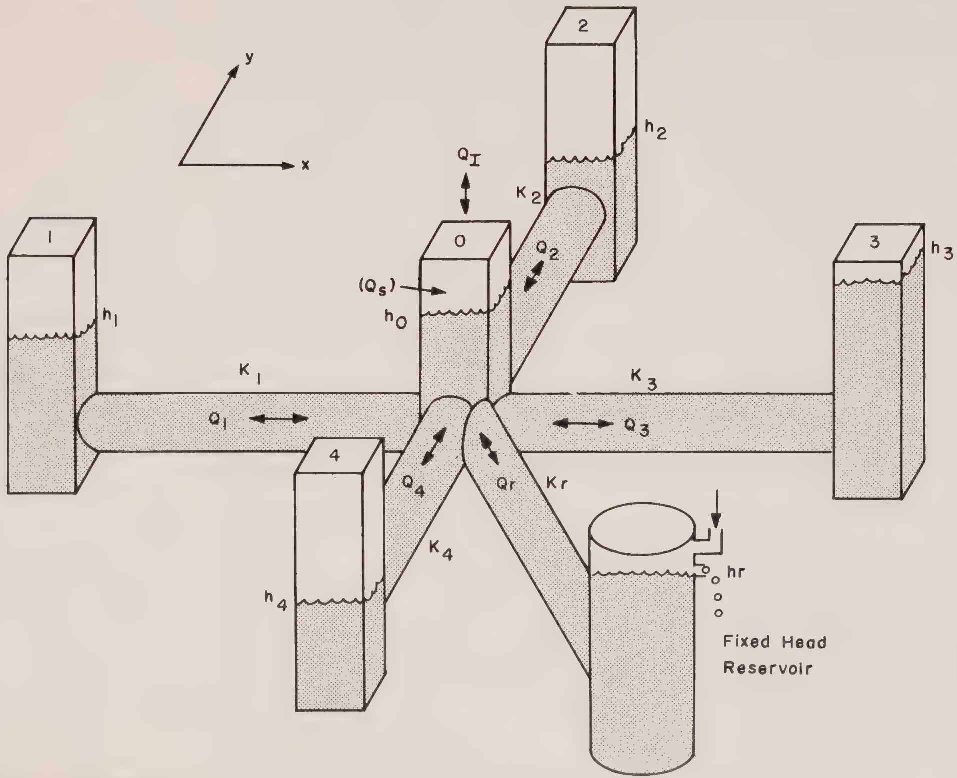


Figure 28. Node network and transmissivity values used for the ground water model.



LEGEND

h_0, h_1, h_2, h_3 and h_4	Piezometric heads at nodes 0,1,2,3 and 4.
Q_1, Q_2, Q_3 and Q_4	Flow between nodes 1,2,3 and 4 and central node.
K_1, K_2, K_3 and K_4	Effective permeability from nodes 1,2,3 and 4 to central node.
hr, Qr and Kr	River head, flow to or from the river, and effective permeability of river bed.
Q_I	Recharge (or pumping, if negative) flow to central node.
Qs	Water released from or added to storage. This quantity will be zero for steady state conditions.

Figure 29. Diagrammatic representation of the water budget around a node.

method of solution for the equations (Southwell, 1946) where the head at the central node ("O" in Figure 29) is overcorrected according to the heads at the adjacent nodes. For the dynamic case, the solution is more efficiently found by using an alternating direction implicit method of solution (Peaceman and Rockford, 1955), where the heads along any row or column are found by consideration of the heads at adjacent rows or columns along with the pumping and storage stresses. These techniques have been written into two computer programs that serve to solve the steady state and dynamic ground water problems. (Pricket and Lonnquist, 1971).

Model Testing

A number of tests were carried out in the models to verify the programs. A number of runs were made with both the steady state and the dynamic model on simple well problems, where the drawdowns could be calculated theoretically. The programs were found to model these conditions accurately.

Using the initial values for the Blue Springs Creek data, a sensitivity analysis was also carried out. The effect on the output of a change in any of the input values was tested in a qualitive manner. Table 10 shows the effect of an increase in the transmissivity, recharge, storage coefficient or river permeability on the output of the model. The results from these tests were used during the later optimization runs for the Blue Springs Creek basin.

PREPARATION FOR THE MODEL

To perform steady state modelling, the following data are required at each node:

a) transmissivity (IGPD/ft),

b) recharge (IGPD/ft^2) or pumpage (IGPD),

c) river head elevation (ft) and the permeability between the river bottom and the aquifer (IGPD/ft^2),

d) heads (ft) at each node for verification of the model.

For the dynamic model, further data are required:

e) storage coefficient (Imp. gal/ft^3), and the recharge rate each month, and the heads at observation wells for each month.

The model was run using the above data and the simulated and the measured heads were compared. On the basis of any discrepancies between the known and modelled heads, the input data to the model were adjusted. Initial estimates for the input data were obtained from the hydrogeological calculations carried out in Chapter 3; their uses in the model development are described below.

TABLE 10. RELATIVE EFFECTS OF A CHANGE IN THE INPUT PARAMETERS ON THE MODEL OUTPUT

Increase in values of:	Resultant Change in Output:			
	Steady State Baseflow	Baseflow Fluctuations	Steady State Water Level	Water Level Fluctuations
Transmissivity	None	None	Reduced	Reduced
Recharge	Increased	Increased	Increased	Increased
Storage Coefficient	None	Reduced	None	Reduced
River Permeability	None	Reduced	Reduced	Reduced

Transmissivity

The transmissivity data were derived from the specific capacity data of the wells in the area (Section 3.2). The aquifer was found to be very heterogeneous and wide variations between the transmissivities of nearby wells were often found.

When considering regional ground water flow patterns, the mean transmissivities of different areas in the basin are important. Thus the modelled area around the Blue Springs Creek basin was divided into several large rectangular areas (Figure 28) and the mean transmissivity in each area was calculated from the wells in that area for initial input to the model.

The presence of solutional channels in a karst area can upset the calculated transmissivities and the flow patterns. In the Blue Springs Creek basin the karst was not considered to be well enough developed to seriously perturb the transmissivity values used in the model.

Recharge to the Ground Water

Estimates of the ground water recharge were determined in two ways: from considerations of the mean baseflow to determine the long term mean recharge for input to the steady state model, and by examining the water balance in the basin to determine the time varying recharge rates for input to the dynamic modelling.

The mean baseflow from the Blue Springs Creek basin above the federal gauge, for the three years 1966 to 1969, was determined to be 14.0 cfs (Section 3.4.1). The mean recharge was assumed to be equal to the long term mean baseflow; this figure was used as the steady state model input.

The effect on the aquifer of pumpage from wells was determined. There are about 200 recorded water wells in the Blue Springs Creek basin, all of which are farm or domestic wells, many of which are now abandoned. Assuming that all of these wells were pumped at an average of 200 IGPD, the total pumpage would be less than four per cent of the baseflow determined above. For this reason the pumpage effect of the wells on the aquifer system was considered to be negligible.

The time varying recharge rates were calculated by considering the

water budget in the basin, as described in Section 3.3.1. In this section the recharge was calculated by considering the ground water budget. The monthly recharge figures used are shown in Figure 13.

River Heads and Permeabilities

In the Blue Springs Creek basin all of the water entering the ground water was assumed to ultimately discharge into the streams as baseflow. For the model input, stream elevations and the permeability between the streams and the aquifer were determined.

The elevations of the streams in the basin and the surrounding area were determined from topographic maps. As the stream course rarely coincided with any node in the superimposed model grid network, the stream system had to be "squared up" to run through the nodes, as shown in Figure 30. The stream elevations at the points nearest to the river nodes were then determined from the maps.

The streambed leakage factor between the river and the aquifer is difficult to determine. The streambed leakage factor at a node can be given as:

$$SL = \frac{Plw}{mA} \tag{16}$$

where SL is the streambed leakage factor at the node,

P is the mean permeability of the river bed material,

l is the length of river associated with the node,

w is the mean river width,

m is the average thickness of the river bed material,

and A is the area associated with the node.

The thickness and permeability of the river bed material are difficult to determine. Therefore, Equation 16 could only be used to derive a general index for the streambed leakage factor. After some tests with the model, an effective permeability of 2 IGPD/ft^2 was assigned to the rivers, 1.0 IGPD/ft^2 to large creeks, 0.5 IGPD/ft^2 to medium creeks and 0.1 IGPD/ft^2 to small creeks. These permeabilities represent very good hydraulic conductivity between the streams and the aquifer, and effectively render the heads in the aquifer at the stream nodes similar to the elevations in the streams.

Piezometric Heads

The heads in the aquifer, which were required for model verification, were obtained from the water well information. The long term mean heads at each node inside the basin were derived from the piezometric surface map (Map 7). Data on the change in elevation of the water at six observation wells were extracted on a monthly basis and used for verification of the dynamic model.

Storage Coefficients

The storage coefficient was estimated at about $2\text{x}10^{-5}$ from pumping

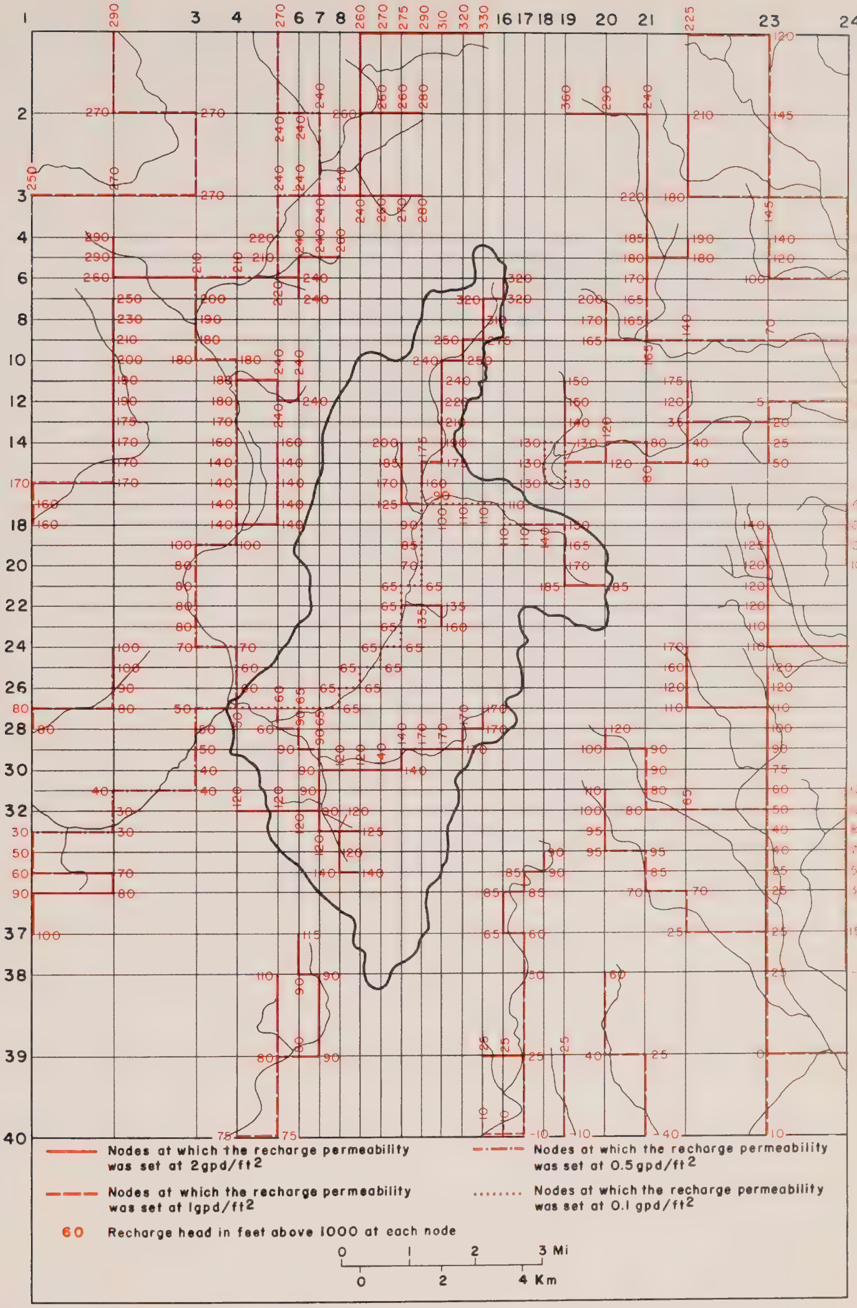


Figure 30. Modelled stream system heads and permeabilities in the Blue Springs Creek basin.

tests on wells near Rockwood (Section 3.2.1). This figure tends to indicate confined conditions. The storage coefficient was also estimated at 4×10^{-2} from water balance considerations explained in Section 3.4. This latter estimation indicates water table conditions. The model was run using both estimates for the storage coefficient.

STEADY STATE MODEL RESULTS

During initial runs of the model, it was found that the variations in the heads over the basin were larger than the measured observed variations. This difference could be corrected by increasing the transmissivity values used in the model, or by decreasing the recharge rate over the whole basin (Table 10). As the baseflow from the model is proportional to the infiltration rate, and was found to be approximately correct during the initial runs, the variations in head were consequently subdued by increasing the model transmissivities. It was found that increasing the transmissivities by 1.5 times their calculated values subdued the heads sufficiently over a large part of the basin. The higher transmissivities in the model, compared to the values found from the wells, are concluded to be due to the nature of the aquifer. In many consolidated rocks the majority of the flow takes place in joints, fissures and bedding planes. Many wells may penetrate poorly fissured portions of the rock and so record a lower transmissivity value than the true regional value. It appears that the regional transmissivity values in Blue Springs Creek basin are therefore about 1.5 times the median values found from the wells in the area.

After adjusting the transmissivity values in this way, errors were still found in the modelled heads near Blue Springs Creek, where the heads were too subdued, and in the northern and southern part of the basin, where the modelled heads were too high. These discrepancies could be corrected by adjusting either or both of the modelled transmissivities or recharge rates near the creek and in the southern and northern portions of the basin. It was found that the baseflow at the Cedar Grove gauge (located in Figure 15), which drains the northern part of the basin, was smaller (5.3 inches a year) than the baseflow at the federal gauge (9.8 inches a year), which drains the central and northern part of the basin, indicating that the discharge rate is less in the Cedar Grove sub-basin than in the basin as a whole. The errors in the modelled heads were therefore corrected by adjusting the recharge rates, while leaving the transmissivities at their previous values. As the Cedar Grove sub-basin generally has at least 50 feet of overburden above the dolomite, the recharge was decreased to 7 inches per year in the regions having more than 50 feet of overburden above the aquifer (Map 5). To maintain the measured baseflow from the federal gauge sub-basin, the recharge was increased to 15 inches per year over the exposed dolomite, where the infiltration capacity is higher than in regions covered by thick till. The final recharge rates used are shown in Figure 31.

Although adjusting the recharge rates over the basin gave a satisfactory head distribution from the model, consideration of the relatively low runoff rate from the Cedar Grove sub-basin (Table 7) suggests that there may be significant ground water flow from the northern to the southerly parts of the basin. Such a flow is consistent with the piezometric information, and would be possible in karst areas, but further work would be needed to prove that such a flow exists. This could be modelled by including a large anisotropic transmissivity in the

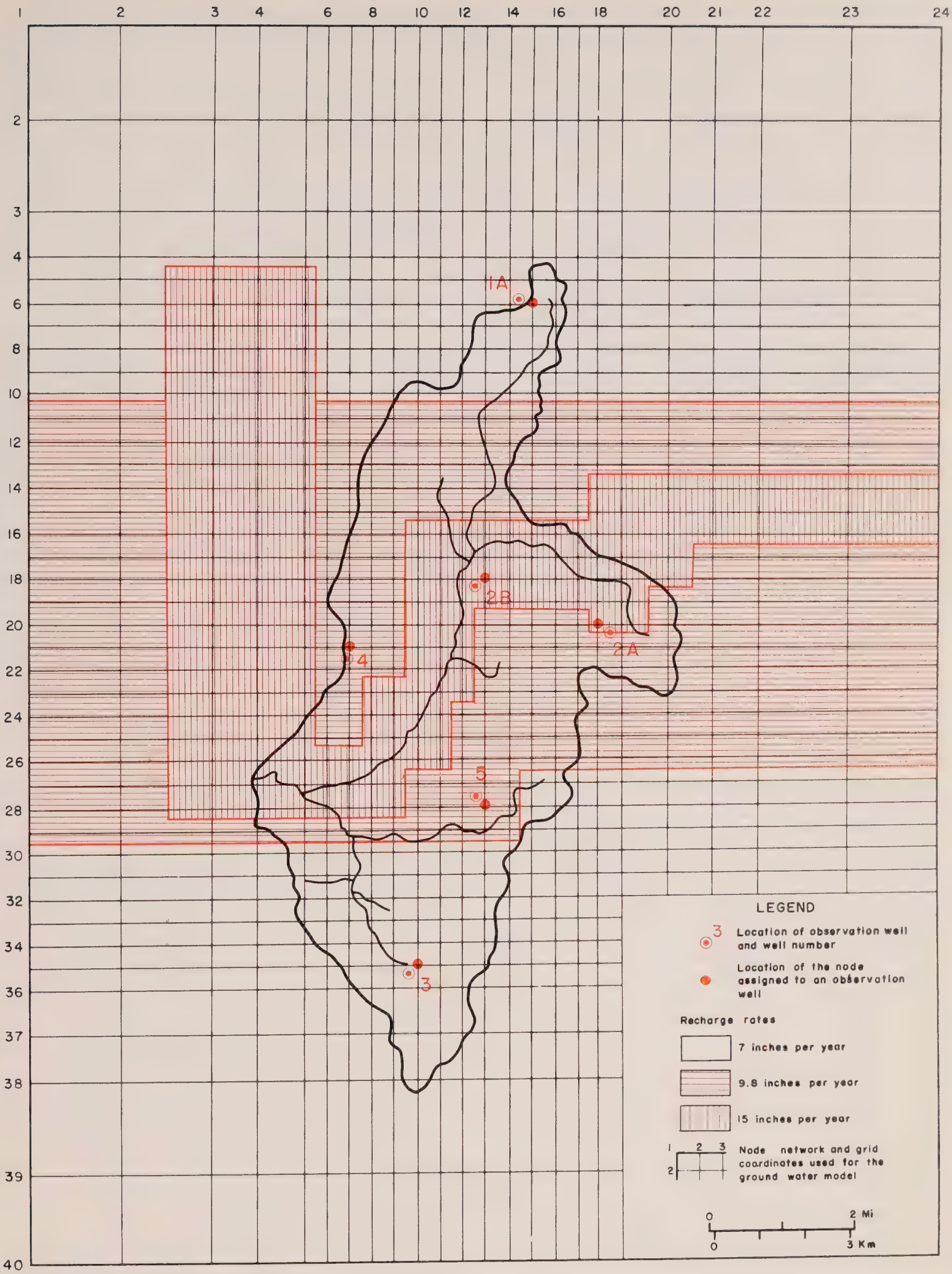


Figure 31. Long term mean effective recharge rates and location of the observation well nodes.

north-south direction, to allow significant flow to occur from north to south. However, such a model was not tried, as adjustment of the recharge rate over the basin gave satisfactory results.

Using the adjusted transmissivity and recharge figures, the modelled heads and a histogram of the differences between the modelled and measured heads for the nodes inside the basin, are shown in Figure 32. At the majority of the nodes (approximately 70 per cent), the differences were less than 10 feet, and the largest difference was 28 feet. The modelled average baseflow in the federal gauge sub-basin was 9.79 inches per year, compared to the measured baseflow of 9.8 inches per year.

DYNAMIC MODEL RESULTS

The dynamic ground water model was run to predict the monthly variations in the heads throughout the area. This model used the transmissivity and mean recharge values determined in the steady state model and used the head distribution found from the steady state model as the initial heads. The dynamic model was used to estimate the storage coefficients throughout the basin, and to verify the variations with time of the recharge rates.

The dynamic model was run for six years from October 1966 until September 1972, using a one month time step, and the calculated recharge rates (Figure 13). The model was also run for an additional twelve time steps using the long term mean infiltration rate to allow the heads to settle to the mean values before the real data were inserted.

During initial runs of the model, using the small storage coefficient representing confined conditions, the modelled head variations at all the observation wells were considerably larger than those measured. The storage coefficients had to be increased to moderate the modelled head variations. The modelled head variations at the six observation wells were brought close to the measured heads by changing the storage coefficient around each well. Due to the small number of observation wells in the basin, the storage coefficients had to be changed over large areas around each well (Figure 33). The higher storage coefficients (from 0.04 to 0.16) used are consistent with water table conditions, and are similar to the values derived by consideration of the ground water balance in Section 3.3.1. The storage coefficients derived from the pumping tests were not typical of the conditions over the majority of the basin.

The modelled heads, using these modified storage coefficient figures, are compared with the measured heads in the six observation wells in figures 34 to 39. As the true elevations of the observation wells were not exactly known, and as a positional error was introduced when the wells were assigned to a node position, the elevations of the average modelled and average measured heads at the observation wells were somewhat different (Table 11). However, the variations in the modelled and measured heads generally agree fairly well at the six observation wells (figures 34-39). The standard deviations of the modelled and measured heads, listed in Table 11, agree closely at the six wells, and the mean error between the head variations varies from 0.53 feet at well 2B to 1.64 feet at well 4 (Table 11).

The measured and modelled baseflows are compared in Figure 40. The baseflows generated by the model are the deep or slow responding ground water discharges (shown diagrammatically in Figure 21) derived from the

ground water reservoir extending over the whole basin. The measured baseflows, however, include the same deep or slow responding discharges as well as the shallow or fast responding discharges, some of which may be derived from local perched ground water reservoirs. In addition, a certain amount of interflow may be included in the measured baseflow figures, due to the method of analysis (Section 3.4.1). For this reason, the modelled baseflows are less variable than the measured values as shown in Figure 40.

CONCLUSIONS

The steady state model was found to simulate the measured heads in the basin closely with the average error between the measured and modelled heads being 6.5 feet and the largest error 28 feet. The total relief of the piezometric surface in the basin was 260 feet. In addition, it was postulated from the steady state model that the regional transmissivities were about 1.5 times the mean transmissivities found from the water wells, and that the infiltration varied from 7 inches a year over the areas covered by more than 50 feet of till, to 15 inches a year over the areas near the creek where the till was thin or absent.

The dynamic model simulated the head variations in the observation wells reasonably well. The storage coefficients around the wells were estimated by the requirements of the model response, and ranged from 0.04 to 0.16 in different areas of the basin. The model simulated the deep seated baseflows, which were considerably less variable than the estimated baseflows.

TABLE 11. COMPARISON OF THE MEAN WATER LEVELS AND WATER LEVELS FLUCTUATIONS IN SIX OBSERVATION WELLS

Well No.	Elevation of Well Head (feet)	Measured Mean Depth to Water (feet)	Modelled Mean Water Level Elevation (feet)	Difference in Mean Water Level Elevations (feet)	Mean Error Between the Measured and Modelled Head Variations (feet)	Standard deviation of the water level fluctuations	
						Measured	Modelled
BS-1A	1350	45.3	1297.6	7	1.060	1.97	1.97
BS-2A	1200	5.5	1167.5	27	1.261	1.54	1.39
BS-2B	1120	29.2	1096.2	-5	0.532	0.92	0.91
BS-3	1150	8.4	1134.2	7	0.624	1.08	1.15
BS-4	1170	12.0	1131.9	26	1.638	2.62	2.63
BS-5	1190	13.4	1155.8	21	0.727	1.67	1.57

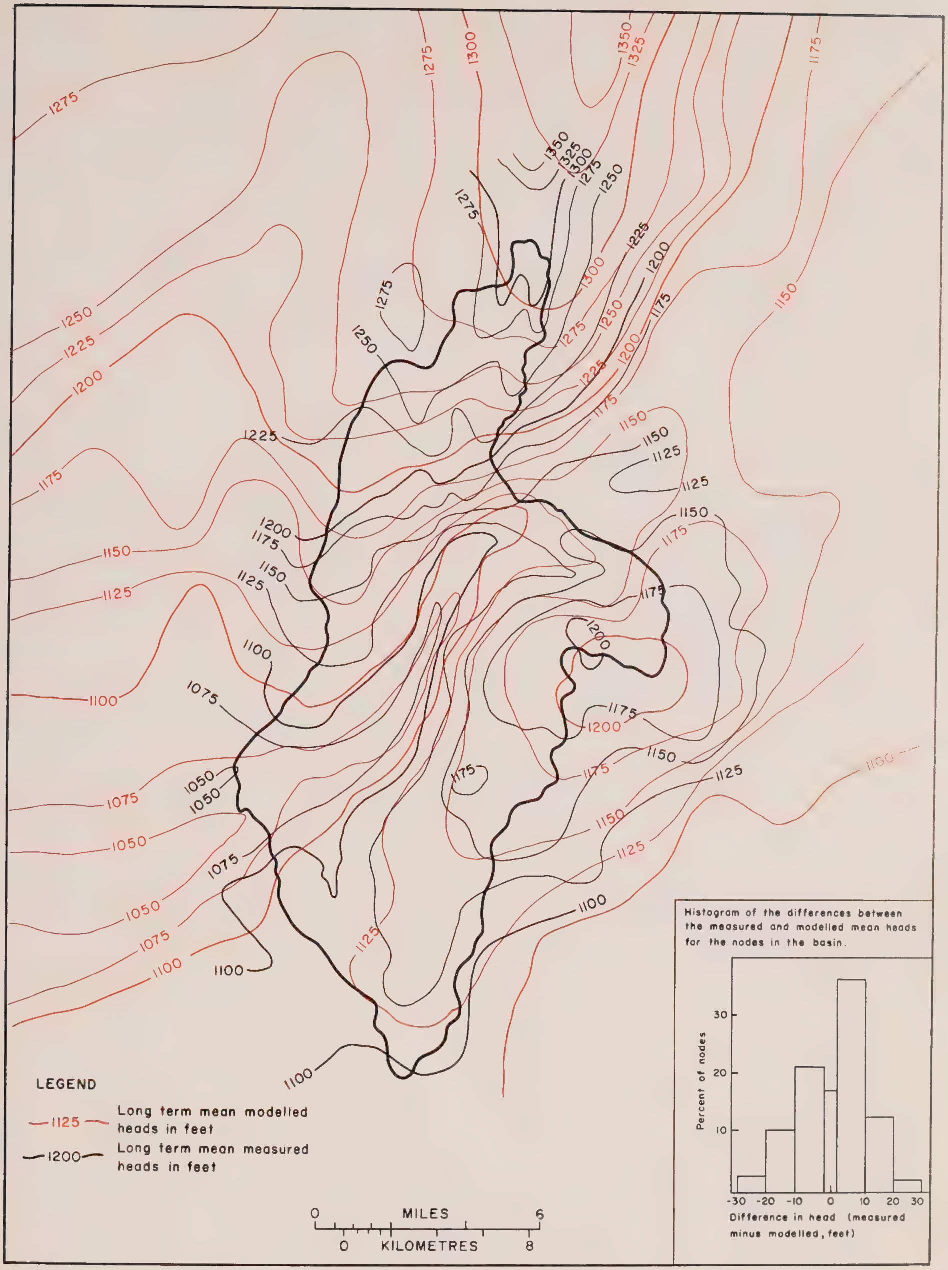


Figure 32. Comparison of the measured and modelled steady state heads in the Blue Springs Creek basin.

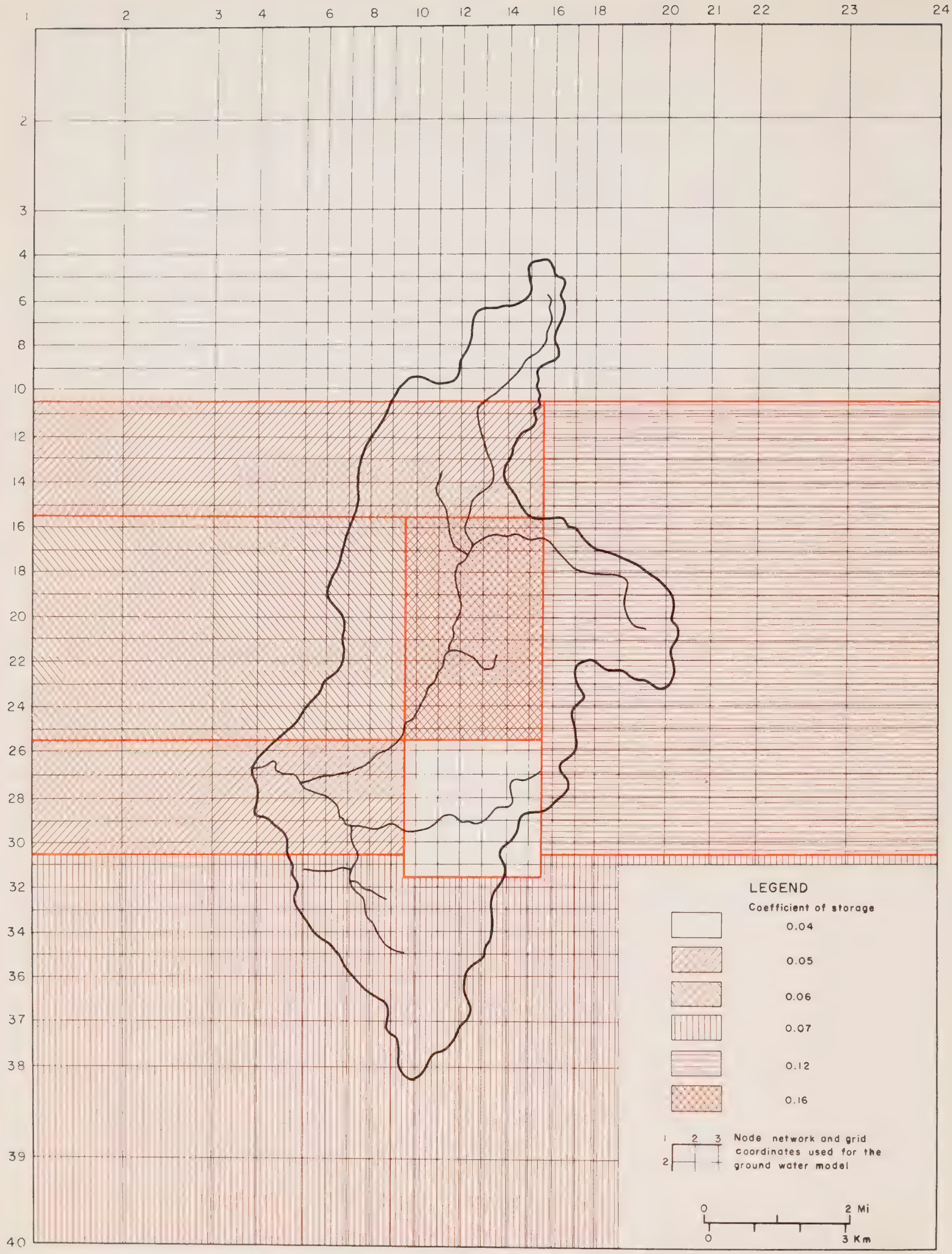


Figure 33. Coefficient of storage values used in the model.

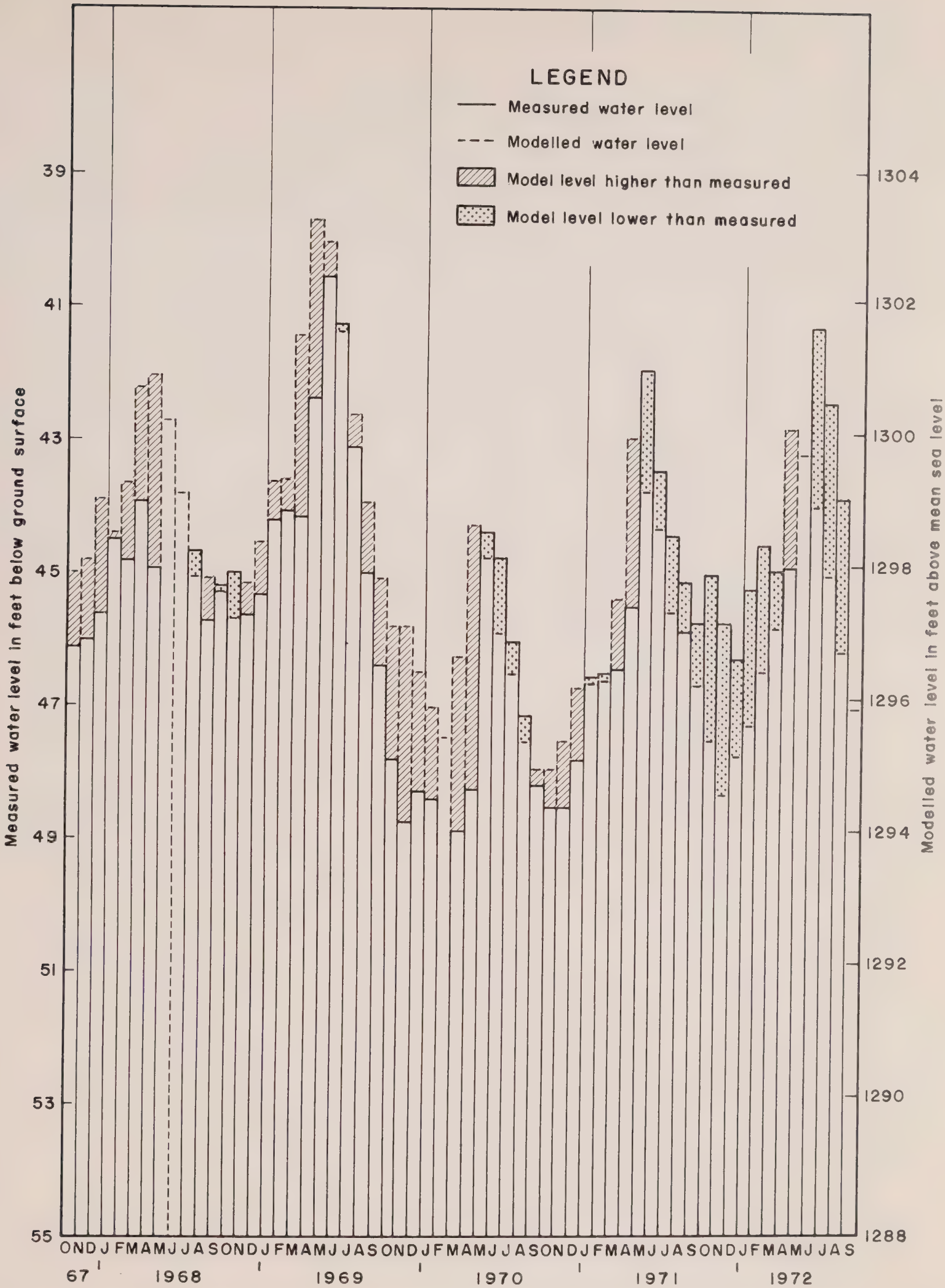


Figure 34. Modelled and measured heads at observation well BS - 1A.

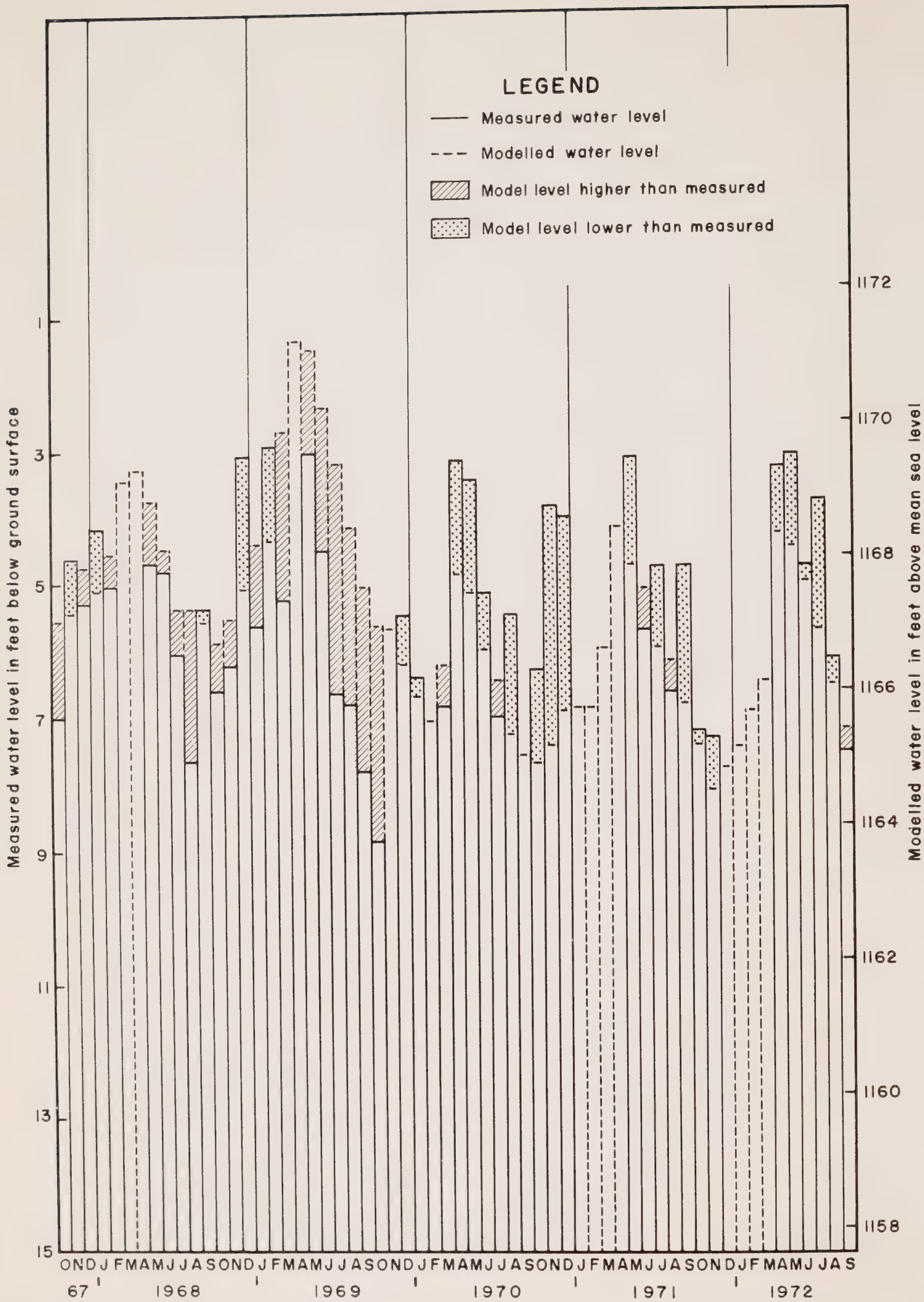


Figure 35. Modelled and measured heads at observation well BS - 2A.

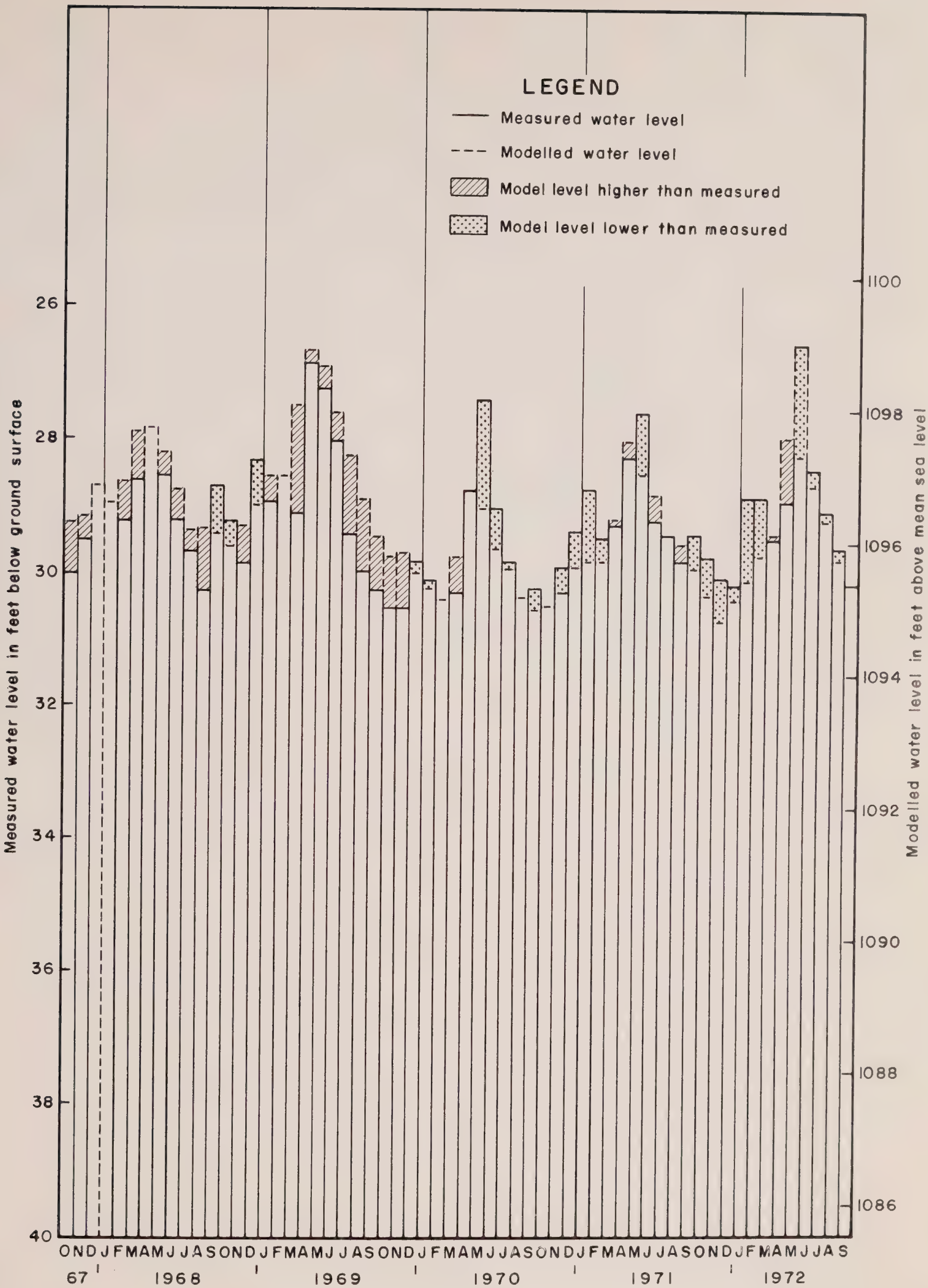


Figure 36. Modelled and measured heads at observation well BS – 2B.

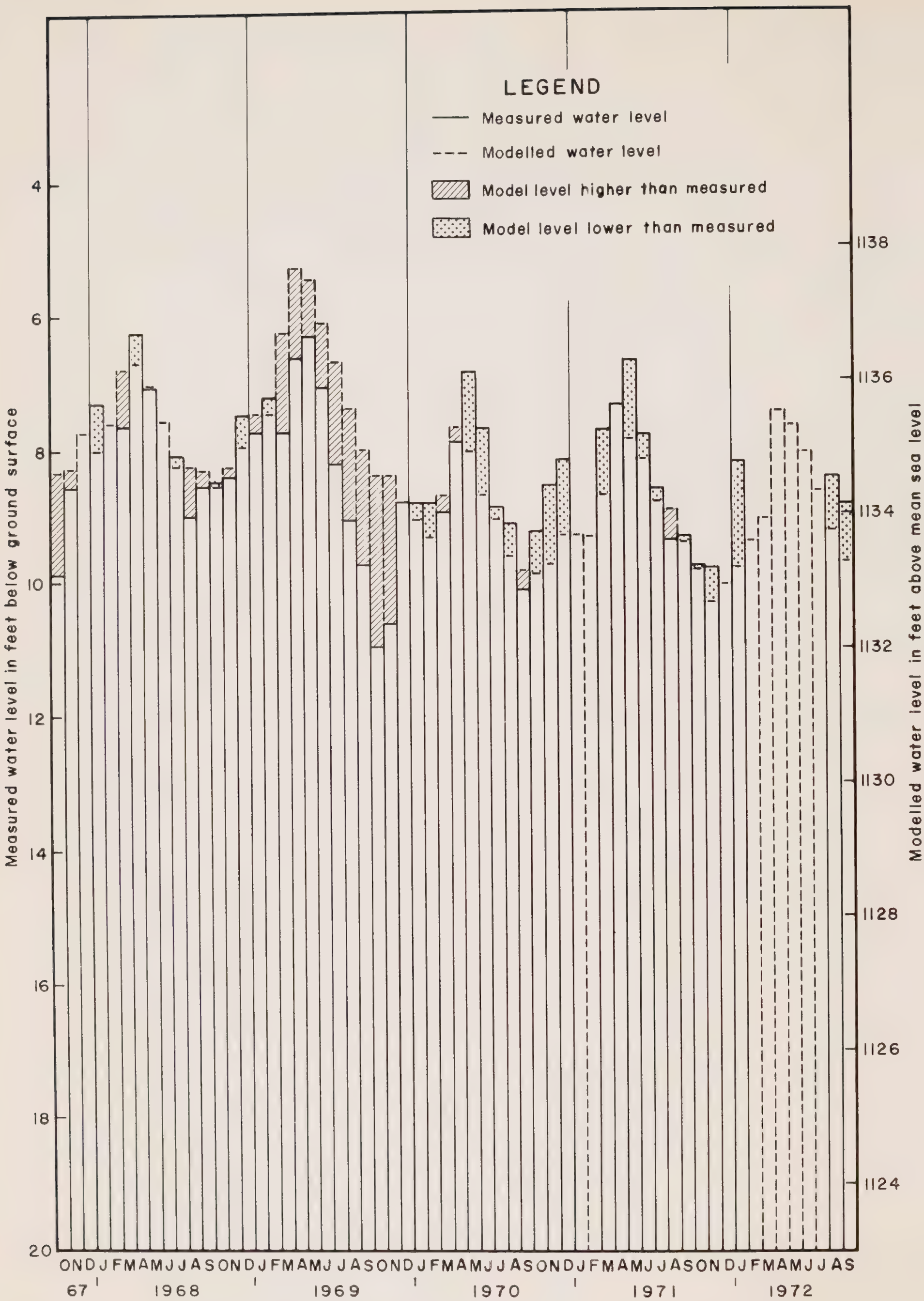


Figure 37. Modelled and measured heads at observation well BS – 3.

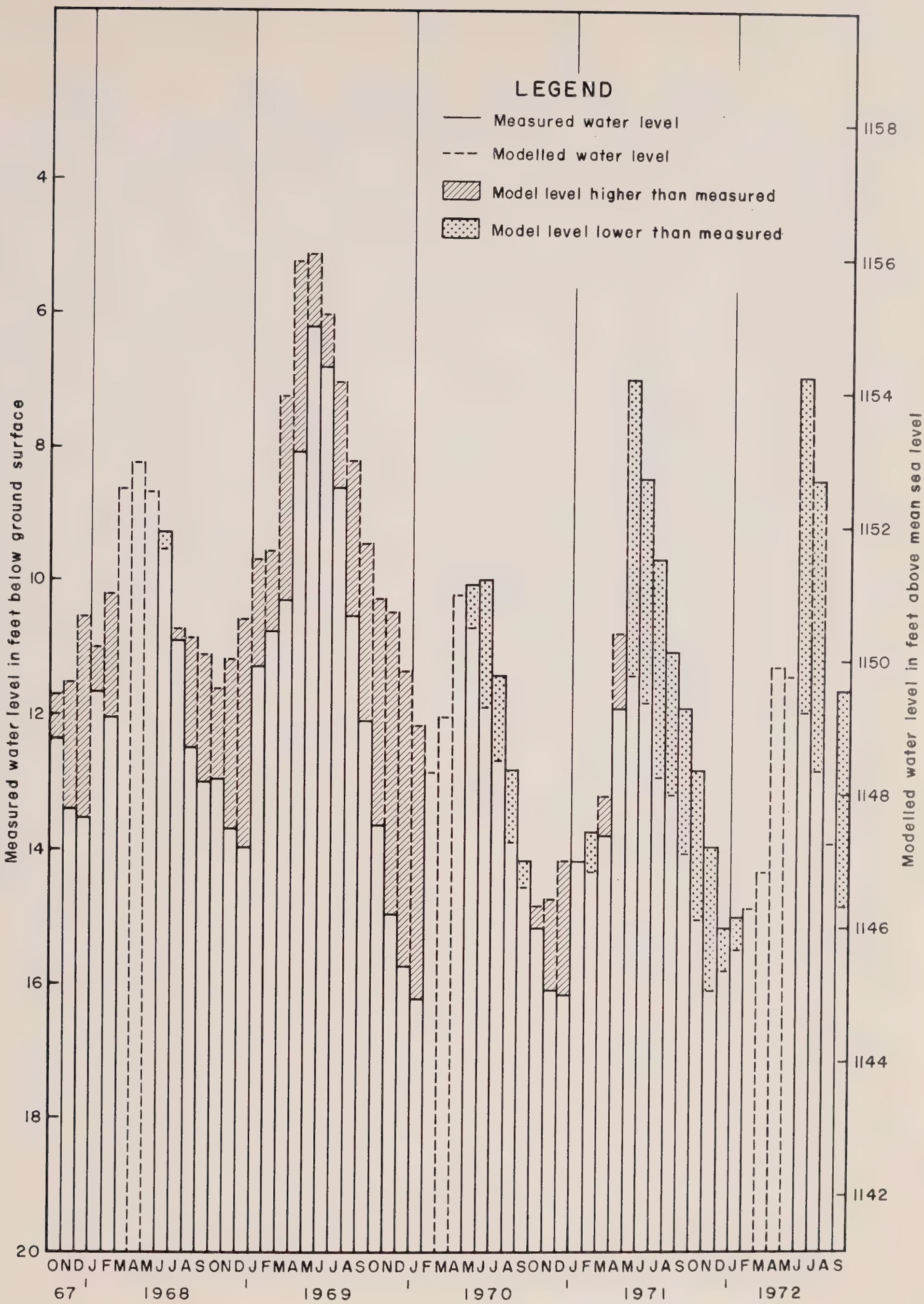


Figure 38. Modelled and measured heads at observation well BS-4A.

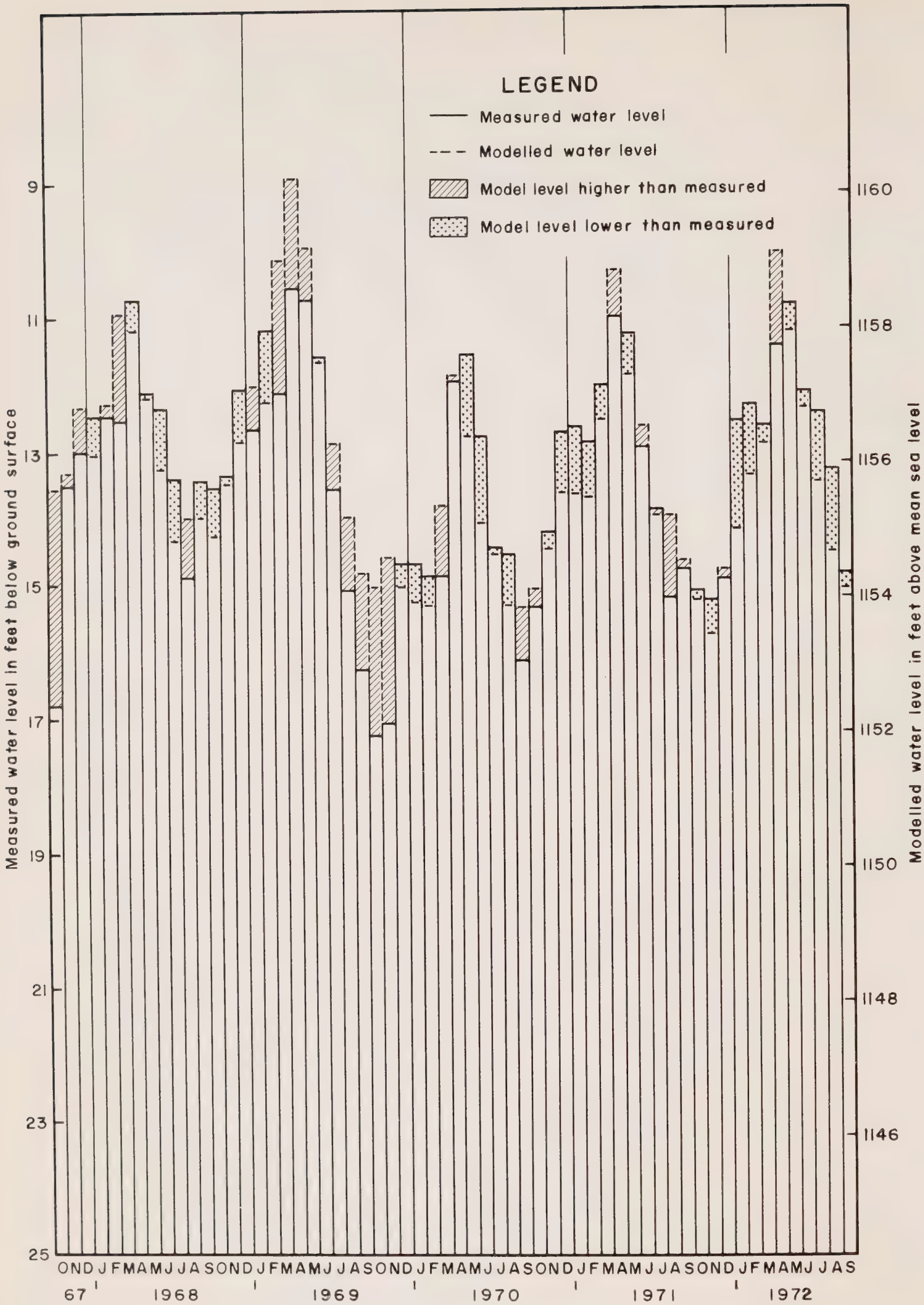


Figure 39. Modelled and measured heads at observation well BS – 5.

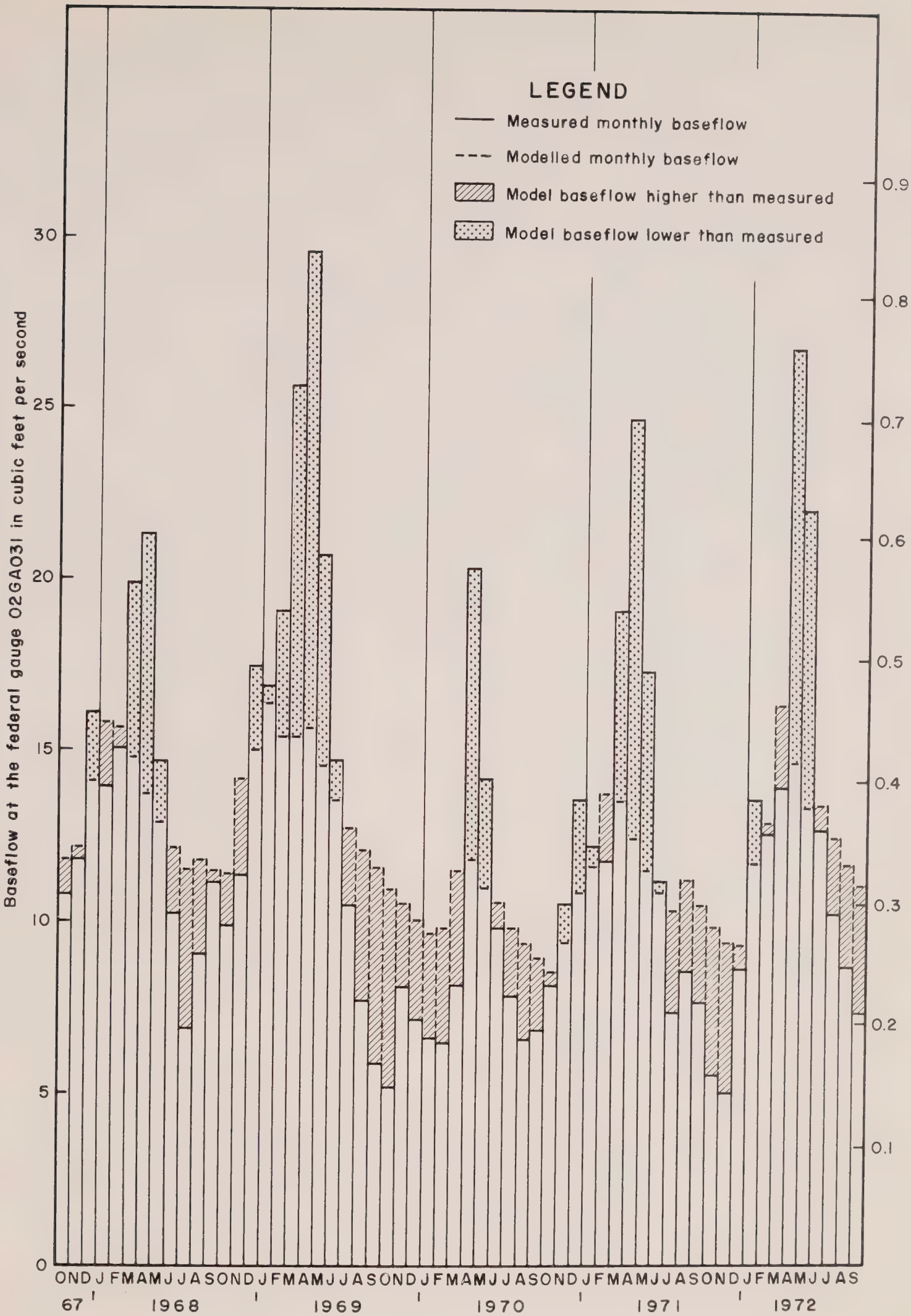


Figure 40. Modelled and measured baseflow values for the federal gauge sub – basin.

HYDROCHEMISTRY

Chemical composition of water in a river basin is an important consideration in any hydrological study. The suitability of the ground or surface waters for use in industry, irrigation or for drinking purposes can be assessed, and ground water flow patterns can be estimated or confirmed by a study of the hydrochemistry. In karst basins where the landforms are largely developed by solution of the local rock strata, an analysis of the hydrochemistry is vital to an understanding of the karst processes. In particular, the relative saturations of the various minerals in the water are needed in order to assess the areas where solution of the rock is likely to take place. In this study, the hydrochemistry of the water was used to confirm the generalized ground water flow patterns and to study the saturations of the water with regard to dolomite and calcite in order to examine the karst development.

WATER SAMPLING

To investigate the lateral variations of the water chemistry over the Blue Springs Creek basin, 73 water samples were collected throughout the basin in July, 1971. Sixty-two samples were collected from existing bedrock wells and 11 samples were taken from springs and streams in and near the basin. The sample locations are shown in Figure 41. As there is only a small number of overburden wells in the basin (see Map 3), detailed sampling of the overburden aquifer was not possible. It is assumed, however, that the water samples taken from the overburden springs in the area, represent hydrochemical conditions in the overburden aquifer itself.

Samples of the well water were obtained from private homes using existing pumps and plumbing facilities to retrieve the samples from the wells. The water was allowed to flow for some time before a sample was taken to allow the stagnant water samples were stored in sealed glass bottles and transferred to the laboratory of the Ministry of the Environment, where the chemical analyses were carried out. The ions analyzed included: calcium, magnesium, sodium, potassium, chloride, sulphate, bicarbonate, nitrate and iron; pH and conductivity were also measured. The ionic concentrations expressed as milli-moles/litre (Hem, 1970) for the 73 water samples are given in Appendix 2.

POTABILITY OF THE WATER

Drinking water should not contain constituents in large enough concentrations to render it harmful, unpleasant to drink or unsuitable for household purposes. Limits on the concentrations of constituents have been listed by the Ministry of the Environment (1974). The published criteria for private water supply are divided into two categories, the permissible and the desirable criteria. Water containing constituents which generally exceed the permissible limits is not fully acceptable for water supply. Water that meets the desirable limits, which are more stringent than the permissible limits, provides good quality water supplies. The permissible and desirable criteria for the constituents that were analyzed for in this study are shown in Table 12.

It was found that many of the water samples did not meet the permissible limits for total dissolved solids or for iron. Their locations are shown in Figure 42. Out of a total of 73 samples, 34 had total

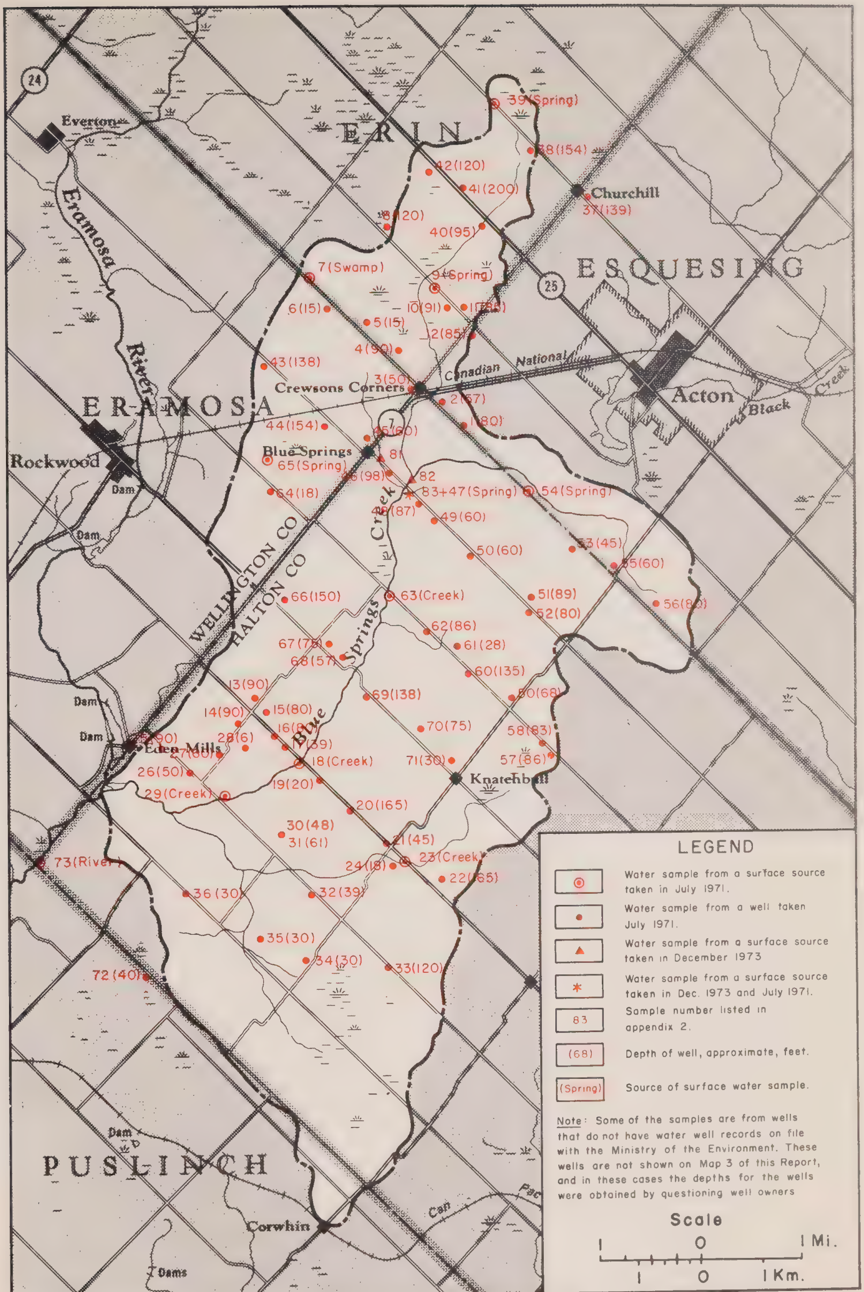


Figure 41. Location of water samples in the Blue Springs Creek basin.

TABLE 12. RECOMMENDED MAXIMUM PERMISSIBLE AND DESIRABLE CONCENTRATIONS OF SOME CONSTITUENTS FOR POTABLE WATER SUPPLIES, AFTER THE MINISTRY OF ENVIRONMENT (1974)

	Permissible criteria (mg/l)	Permissible criteria (milli-moles/l)	Desirable criteria (mg/l)	Desirable criteria (milli-moles/l)
Chloride	250	7.1	<25	<0.71
Iron	0.3	-	virtually	absent
Nitrate (as N)	10	0.72	virtually	absent
Sulphate	250	2.6	<50	<0.52
Total Dissolved Solids	500	-	<200	-
pH	6.0-8.5 units			

TABLE 13. Classification of Hydrochemical Facies, after Back (1960)

	Percentage of constituents, in equivalents per million			
	Ca+Mg	Na+K	HCO_3	$Cl + SO_4 + NO_3$
Cation facies:				
Calcium-magnesium.....	90-100	0-10	....	....
Calcium-sodium........	50-90	10-50	....	....
Sodium-calcium........	10-50	50-90	....	....
Sodium-potassium......	0-10	90-100	....	....
Anion facies:				
Bicarbonate...........	.....	......	90-100	0-10
Bicarbonate-chloride-sulfate.............	.....	......	50-90	10-50
Chloride-sulfate-bicarbonate.........	.....	......	10-50	50-90
Chloride-sulfate.....	.....	......	0-10	90-100

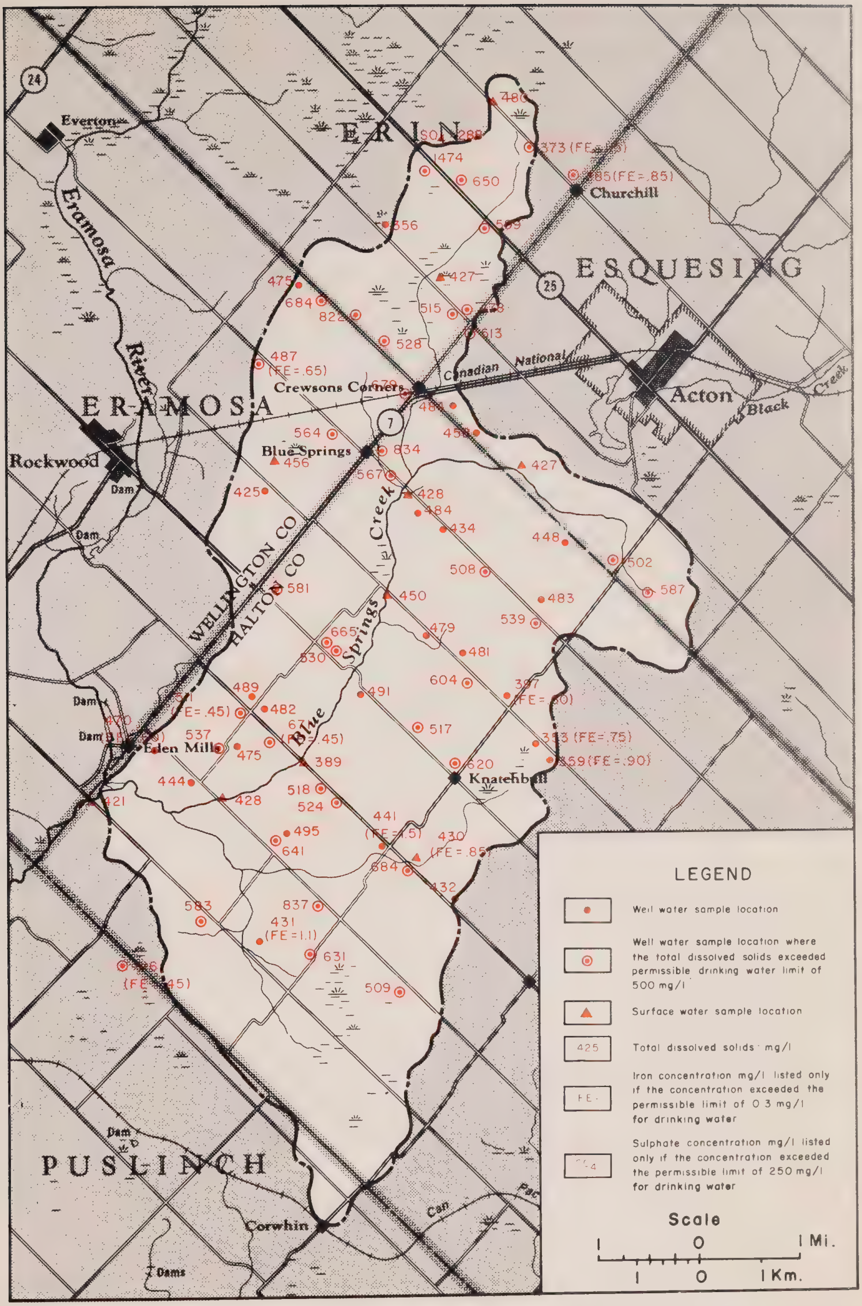


Figure 42. Chemical potability of water samples in the Blue Springs Creek basin.

dissolved solids above the limit of 500 mg/l. Thirteen samples had an iron concentration above the 0.3 mg/l limit, although it is possible that the high iron content in some samples could be from the pumps and plumbing systems. In general, however, water can be treated readily in home systems to remove excess iron.

Over half of the waters sampled in the Blue Springs Creek basin had one or more constituents which were above the permissible limits.

Hardness of the Water

Hardness is a measure of the amount of calcium and magnesium in solution. Hard water is objectionable because it tends to form scale in water heaters, pipes, kettles and boilers, and because a large amount of soap is needed to form a lather. Water hardness is often classified from soft (less than 60 mg/l equivalent $CaCO_3$) to very hard (over 180 mg/l equivalent $CaCO_3$), (Dunfor and Becker, 1964). All the water samples in the Blue Springs Creek basin (except sample 60, which is discussed in Section 5.3) were very hard (Appendix 2), containing from 216 to 600 mg/l of equivalent $CaCO_3$. The majority of this hardness is of the temporary type (e.g. carbonate hardness) and could be reduced by suitable home treatment systems.

GENERAL CLASSIFICATION OF THE WATER

In order to assess the differences in the chemistry of the surface and ground waters in the basin, the water samples were classified into hydrochemical facies using the terminology developed by Back (1960). This classification is based on the dominant anions and cations present in the water. The various hydrochemical facies are listed in Table 13 and have been inserted on the hydrochemical facies diagram, Figure 43. Generally, a natural water will range from a predominantly bicarbonate water, through a sulphate water, to a predominantly chloride water as the degree of 'metamorphism' of the water increases (Chebotarev, 1955).

The percentages of milli-equivalents per litre (meq/l) of the various cations and anions were plotted on the trilinear and parallelogram diagrams shown in Figure 43. It can be seen from Figure 43 that the majority of the water samples were calcium-magnesium, and bicarbonate-chloride-sulphate facies. Most of the water is therefore in its first or second phase of 'metamorphism'.

Sample 42 is unusual in that the potassium concentration is very high. Potassium rarely exceeds the sodium concentration (Hem, 1970), but in sample 42 the potassium concentration is over five times the sodium concentration. The chloride, sulphate and nitrate concentrations are also high, leading one to suspect that this well is being contaminated by fertilizers, sewage or farmyard wastes.

Areal Variations in the Degree of Metamorphism of the Water

The lateral variations in the degree of metamorphism were investigated. The chloride and sulphate to bicarbonate ratios were plotted for the basin as shown in Figure 44 (in pocket), where the high numbers indicate a relatively high degree of metamorphism.

There are some general trends in Figure 44, the predominant one being the banding roughly parallel to the creek in the middle and

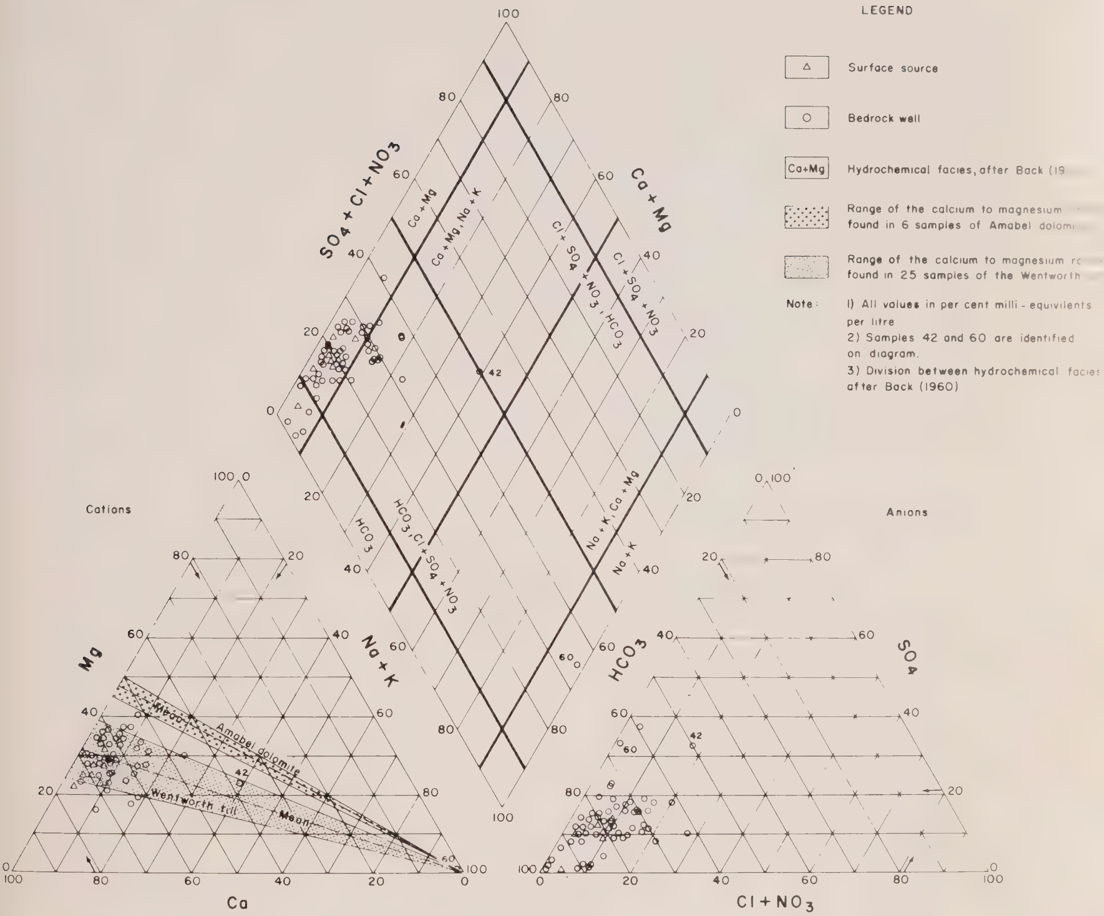


Figure 43. Chemical composition of water in the Blue Springs Creek basin.

southerly parts of the basin. The degree of metamorphism appears low in the ground water recharge and discharge areas and higher in the areas between. In a general way the degree of metamorphism over the basin fits the ground water divide (on Map 7) more closely than the surface water divide.

DEGREE OF SATURATION OF WATER WITH RESPECT TO DOLOMITE AND CALCITE

Chemical solution of the aquifer material will play an important part in the erosion and geomorphic development of any areas underlain by carbonate rocks. In karst areas in particular, the chemical erosion of the rocks predominates over mechanical erosion. The solution of the rocks by the water is limited by the solubility of the aquifer material in the water. Thus an understanding of the solubility of the aquifer material is important for the understanding of the geomorphic development of carbonate areas, and to assess regions where solution of the aquifer material is likely to be significant.

Water which is in contact with soils and carbonate rocks will dissolve some of the minerals. Given time, the solution of the minerals will continue until the water becomes saturated with respect to that mineral. When this occurs, the solubility equation is applicable; for dolomite this can be expressed as:

$$K_D = \frac{\{Ca^{++}\}\ \{Mg^{++}\}\ \{CO_3^{--}\}^2}{\{Ca\ Mg\ (CO_3)_{2_S}\}} \qquad (17)$$

where $\{Ca^{++}\}$ is the activity of the calcium in solution;

$\{Mg^{++}\}$ is the activity of the magnesium in solution;

$\{CO_3^{--}\}$ is the activity of the carbonate in solution;

$\{Ca\ Mg\ (CO_3)_{2_S}\}$ is the activity of the solid dolomite (which by convention is unity);

and K_D is the solubility product for dolomite (which is temperature and pressure dependent).

Equation 17 only applies if the water is in contact and in equilibrium with solid dolomite.

If the water is not saturated with respect to dolomite, the water is considered to be aggressive and is capable of dissolving dolomite. Conversely, the water can be supersaturated, particularly if carbon dioxide is lost, and precipitation of dolomite will tend to occur. A convenient index for the degree of saturation can be given by (Wigley, 1971):

$$SATDOL = \log\ (\{Ca^{++}\}\ \{Mg^{++}\}\ \{CO_3^{--}\}^2) \quad \log\ \{K_D\} \qquad (18)$$

where SATDOL is the saturation index with respect to dolomite: it is positive if the water is supersaturated with respect to dolomite and negative if undersaturated. Comparing equation 17 with Equation 18, it can be seen that SATDOL will be zero if the water is saturated.

A similar index can be given with respect to calcite as:

$$SATCAL = \log (\{Ca^{++}\} \{CO_3^{--}\} - \log (K_c) \qquad (19)$$

where SATCAL is the saturation index with respect to calcite, and K_c is the solubility product for calcite.

In order to determine the saturation indices, the activities of the appropriate ions are needed. The activities and saturation indices can be derived from the chemical analysis figures. The calculations are more complicated if ion pairs are considered, (Garrels and Christ, 1965). Ion pairs have often been neglected in hydrochemistry calculations (for example Van Everdingen 1969), but Wigley (1971) has shown that ion pairs can be significant in hydrochemistry calculations and appreciably alter the calculated ion concentrations and degrees of saturation. In this study, the following ion pairs were considered: $CaSO_4^o$, $MgSO_4^o$, $CaHCO_3^+$, $MgHCO_3^+$, $CaCO_3^o$, $MgCO_3^o$, $NaSO_4^-$, KSO_4^-, $NaCO_3^-$, $NaHCO_3^o$. Other possible ion pairs (for example KCO_3^-, $CaCl^+$) were considered not to be present in significant amounts.

The calculation of the ion and ion pair concentrations was carried out using a computer program which was written and has been described by Wigley (1971). The program calculates the degree of saturation with respect to calcite and dolomite; the results are listed in Appendix 2.

Errors that Affect the Calculated Degree of Saturation of Water with Respect to Calcite and Dolomite

There are a number of errors, which are associated with the sampling and the analysis of water, that may affect the calculated degree of saturation with respect to carbonate minerals. These errors and their effect on the calculated degree of saturation are described in this section.

Certain chemical changes may take place in the water between the sampling and analysis time (Hem, 1970). If the water is supersaturated with respect to any mineral, precipitation will probably take place to bring the water towards equilibrium. Precipitation of calcite or dolomite will decrease the measured amount of calcium, magnesium and alkalinity and will also decrease the measured pH.

If the water is undersaturated, it is possible to lose dissolved carbon dioxide from the water if the partial pressure of the carbon dioxide is above the atmospheric value, particularly if the sample bottles are not completely filled, or if the bottles are left open for a time before analysis. Loss of carbon dioxide will result in a decrease in the alkalinity and an increase in the pH value, and indirectly will result in a reduced degree of calculated undersaturation with respect to calcite or dolomite. If sufficient carbon dioxide is lost, the water may become supersaturated and start to deposit carbonates. To determine the changes in the water, the most sensitive parameter to be examined is the pH: if the pH rises, the water is losing carbon dioxide, and if the

pH falls, carbonates are probably being deposited without loss of carbon dioxide. In many natural waters, the pH rises during sample storage, indicating that carbon dioxide is being lost (Hem, 1970).

To estimate these changes in chemistry that may occur in the water samples from Blue Springs Creek, three samples were collected on December 17, 1973, at the sites marked on Figure 41. The pH and temperature were measured in the field and subsequently analyses were performed in the laboratories of the Ministry of the Environment. The relevant data are shown in Table 14 and Appendix 2. It can be seen that in all three samples the pH did rise somewhat between sampling and laboratory analysis likely due to loss of carbon dioxide from the samples during storage. The calculated partial pressure of carbon dioxide is listed in Table 14 and can be seen to be well above the normal atmospheric partial pressure of 0.03 per cent. The largest pH change occurred in sample 83, which had the highest partial pressure of carbon dioxide and would therefore be most likely to lose the largest amount of carbon dioxide.

The effect of the temperature changes can also be seen from Table 14. The water sampled varied from 7.0 to 8.5°C and the laboratory was assumed to be at 20°C (68°F) during analysis. As water is warmed, carbonates become more soluble and the degree of saturation is reduced. Samples 81 and 82 (which were from surface streams) were both supersaturated with respect to calcite at the time of sampling. When analysed in the laboratory, all samples had equilibriated and were all approximately saturated with respect to calcite. The degrees of saturation with respect to dolomite were generally lower, although samples 81 and 82 were supersaturated at the sampling time.

Generally speaking, the saturations calculated using field pH measurements are below those using laboratory pH measurements. Thus the saturations in Appendix B, which use the laboratory pH and an assumed field water temperature of 10°C, are probably too high. Changes in the degree of calcite and dolomite saturations should, however, be generally less than 0.3 and 0.6 units, respectively.

Under ideal conditions, pH measurements can be accurate to 0.05 units. Thus, as the calcite saturation is approximately proportional to the hydrogen ion concentration, and the dolomite saturation approximately proportional to the square of the hydrogen ion concentration, the corresponding errors in the degree of calcite and dolomite saturations are about 0.05 and 0.1 units, respectively. Thus, an analysis which had a degree of calcite saturation from -0.05 to 0.30 units, or a degree of dolomite saturation from -0.1 to 0.60 units, was considered to be fully saturated.

Areal Saturations

The degree of saturation of the water determines if the water is capable of dissolving the minerals in the aquifer material. If the water is undersaturated, solution can occur. If saturated, the water is in equilibrium with the minerals and no solution or precipitation of the minerals will occur. If the water is supersaturated, which can occur when ground water resurges and loses carbon dioxide to the air, precipitation of minerals is probable.

The degree of saturation of the water with respect to calcite and dolomite was studied in the Blue Springs Creek basin to determine the areas where solution and precipitation are probable. The degree of saturation of the water samples is shown in Figure 45 (in pocket). Most

TABLE 14. A COMPARISON BETWEEN FIELD AND LABORATORY pH AND TEMPERATURE MEASUREMENTS FOR THREE SAMPLES IN BLUE SPRINGS CREEK AND THEIR EFFECT IN SATURATION CALCULATIONS

Sample No	Field Temperature ^{o}C	Free ion concentration** milli moles/litre Ca^{++}	Mg^{++}	HCO_3^-	Laboratory pH	Field pH	Partial Pressure* CO_2, % Atm.
81	7	1.800	0.717	3.624	8.0	7.9	0.21
82	7.5	1.588	0.912	3.808	8.1	8.0	0.20
83	8.5	1.671	1.159	4.541	7.9	7.6	0.59

Sample No	Calculated Saturation with respect to Calcite			Calculated Saturation with respect to Dolomite		
	Using Field pH & Temp.	Using Lab pH & Field Temp	Using Lab pH at $20^{o}C$	Using Field pH & Temp.	Using Lab pH & Field Temp.	Using Lab pH at $20^{o}C$
81	0.27	0.27	0.00	0.06	0.17	-0.40
82	0.23	0.35	0.09	0.26	0.49	-0.06
83	0.06	0.26	0.02	-0.25	0.40	-0.11

* Global mean partial pressure of carbon dioxide in the atmosphere is 0.03% Atm.

** The free ions concentration is less than the total ion concentration listed in Appendix A as some of the constituents form ion pairs, which reduces the free ion concentration.

of the samples (47 out of the total of 73) were saturated with respect to calcite. Only four samples were supersaturated, and 22 samples were undersaturated. Most of the samples (55) were undersaturated with respect to dolomite. Only three samples were supersaturated and 15 were saturated with respect to dolomite, although all the well samples were obtained from dolomitic bedrock wells. The reason for the lack of equilibrium with respect to dolomite is further discussed in Section 5.5.3.

In Figure 45, it can be seen that there are no obvious trends in the degree of saturation with respect to dolomite. The undersaturated samples appear to be randomly distributed. The reason for this effect is explained in Section 5.5.3.

The calcite figures also appear to be fairly random; however, many of the undersaturated samples are near the drainage divide where the water is recharging the aquifer. Conversely, many of the saturated samples are in the centre of the basin, where the ground water has had a long residence time in the aquifer and so has had time to equilibriate with the aquifer material.

When the ground water resurges it generally gives off carbon dioxide, as the partial pressure of carbon dioxide is usually above that of the atmosphere. This increases the degree of saturation of the water and may give water which is supersaturated with respect to calcite and dolomite. In this case the minerals would tend to precipitate out. If the ground water resurges at a bedrock spring the precipitate of the carbonate can lead to a deposit of tufa; such deposits were seen at Rockwood just to the west of the basin. In the Blue Springs Creek basin, however, the only water which is supersaturated with respect to both calcite and dolomite is in the main stream (Figure 45). Any precipitation that occurs would be in the sediments giving a calcareous mud or marl deposit.

CALCIUM TO MAGNESIUM RATIO

The calcium to magnesium ratios (Ca/Mg) of the waters in the Blue Springs Creek basin were used to estimate generalized ground water flow patterns in the bedrock aquifer. In general, ground water will tend to equilibriate with the surrounding rock, and with time, the (Ca/Mg) of the water will approach the (Ca/Mg) of the aquifer material. As the chemical changes are not instantaneous, the water which has had a short residence time in the bedrock aquifer, will have a different (Ca/Mg) than the water which has had a long residence time in the bedrock aquifer. Generally the ground water which is being recharged has had a shorter residence time than the ground water that is being discharged. It was found possible to predict, in a general way, the recharge and discharge areas of the bedrock aquifer by using the calcium to magnesium ratio of the waters. To do this it was first necessary to determine the (Ca/Mg) in the bedrock and overburden materials; this is described in the following subsections.

Calcium to Magnesium Ratio of the Amabel Dolomite

The (Ca/Mg) of the Amabel bedrock is given by rock analysis figures quoted by Hewitt (1964). A mean of six samples from quarries near the towns of Milton and Guelph (the quarry localities are shown on Figure 2)

gave a CaO and MgO composition of 30.71 per cent and 20.19 per cent, respectively by weight. Converted to the mole fraction (a mole is the molecular weight of substance in grams), the calcium/magnesium mole ratio is:

$$\frac{(Ca)}{(Mg)} \text{ Mole} = \frac{30.71}{56.08} \times \frac{40.31}{20.19} = 1.09$$

as the molecular weights of Ca) and Mg) are 56.08 and 40.31, respectively. The extreme values for the seven samples gave a (Ca/Mg) ratio of 1.01 at Guelph to 1.30 at Milton. Cowell and Gregor (1973) found a molar ratio from 1.00 to 1.08 with a mean of 1.03 in five samples of the Colpoy Bay and Wiarton members of the Amabel at Rockwood.

The (Ca/Mg) of pure dolomite ($CaMg(CO_3)_2$) is 1.0. Thus the Amabel is a fairly pure dolomite and contains only a small percentage of calcite.

<u>Calcium to Magnesium Ratio of the Wentworth Till</u>

The carbonate composition of the Wentworth till in the Guelph area is quoted by Karrow (1968). The calcite/dolomite ratio was measured using the method of Dremanis (1962), and the average ratio of 25 samples of Wentworth till was 0.63 by weight. Converted to the mole fraction, the ratio is:

$$\frac{(\text{Calcite})}{(\text{Dolomite})} \text{ mole} = \frac{0.63}{100.09} \times \frac{184.41}{1} = 1.15$$

as the molecular weights of calcite and dolomite are 100.09 and 184.41, respectively. As one mole of dolomite contains one mole of calcium and one mole of magnesium, the (Ca/Mg) of the till will be:

$$\frac{(Ca)}{(Mg)} \text{ mole} = \frac{1.15+1.0}{1.0} = 2.15$$

The extreme values for the (Ca/Mg) of the 25 samples of Wentworth till are 1.59 and 2.96.

<u>Comparison of the Calcium to Magnesium Ratios of the Water and of the Aquifer Material</u>

The molar ratios of the dolomite bedrock and overburden have been added to the trilinear diagram, Figure 43. The majority of the waters have a (Ca/Mg) similar to that of the overburden material, although all the wells were in the dolomite bedrock.

In Pennsylvania, Miesler and Becker (1967) found a close correlation between the (Ca/Mg) of the ground water and the bedrock lithology. Wells in dolomite gave a mean (Ca/Mg) of 1.09, while limestone wells gave a mean ratio of about 4. However, the area in Pennsylvania was not covered with much overburden, while in Blue Springs Creek the bedrock is covered with a considerable thickness of till. The till must be affecting the (Ca/Mg) in the underlying dolomitic aquifer.

In the Blue Springs Creek area the water in the till generally becomes almost saturated with respect to calcite (as can be seen from samples 39 and 7, Appendix 2, which are from springs in the till), but

is undersaturated with respect to the dolomite. When this water enters the dolomite bedrock, solution takes place, but as the dolomite dissolves, the calcite concentration must increase and the water becomes supersaturated and thus precipitates calcite. This calcite deposition hinders dolomite solution and prevents the water from reaching saturation with respect to the dolomite and also from approaching the (Ca/Mg) of the dolomite bedrock.

Areal Variations of the Calcium to Magnesium Ratio of the Water

The (Ca/Mg) of the waters throughout the Blue Springs Creek basin were plotted in Figure 46 (in pocket). The ratios appear higher near the basin divides and become lower towards the centre of the basin. The areas of high (Ca/Mg) represent areas of bedrock recharge and the areas of lower ratio are areas where the water has had a long residence time in the dolomite; they also tend to be areas where the bedrock is discharging water. In the extreme northern part of the basin, the (Ca/Mg) is low, which supports the findings of Section 3.3 that this area is a discharge zone for the bedrock aquifer, in spite of the fact that the area is near the topographic divide.

SUMMARY AND CONCLUSIONS

A hydrogeologic study was undertaken in the Blue Springs Creek basin to evaluate the ground water conditions in terms of distribution, quantity and quality. The basin is representative of karst conditions in Southern Ontario, such as those found along the top of the Niagara Escarpment. A summary of the significant results of the study, and the conclusions derived from these results are given below.

1) The hydrogeologic studies reveal that the materials underlying the Blue Springs Creek basin form a complex hydrologic system. This system comprises a dolomite unit, a sandy till unit and a kame, outwash and alluvial deposit unit. The dolomite unit (Amabel dolomite) forms the principal aquifer in the study area. The kame, outwash and alluvial deposit unit combined with the sandy till forms a minor aquifer.

   Most wells constructed in the dolomite bedrock aquifer should yield adequate quantities of water for domestic farm, commercial and smaller industrial uses, but the water will likely be very hard and have a total dissolved solids content above the 500 mg/l permissible limit recommended for public and private water supplies.

2) Based on geomorphic and hydrogeologic evidence, the development of karst features in the Amabel dolomite is minor. There are several caves and springs at Rockwood and a number of surface sinks and springs are found in the basin which interact with the dolomite aquifer. As the karst is not well developed, it does not play a major role in altering the ground water flow patterns.

3) Conditions during the study period (1967-1971) were found to be slightly wetter than the long term mean conditions. The precipitation at Guelph during the study period averaged 33.7 inches/yr (855 mm/yr), which was 3 per cent higher than the 1941-1970 mean of 32.7 inches/yr (830 mm/yr). The streamflow on the Speed River below Guelph (at the federal gauge 02GA015) averaged 212 cfs (6.0 m^3/s) during the study period, which is 8 per cent higher than the mean flow for the period of record which was 197 cfs (5.6 m^3/s).

4) The transmissivity of the overburden was estimated from well records and the median transmissivity was found to be 3,500 IGPD/ft. (600 mm^2/s). A large range of transmissivity values was found; the 10 and 90 percentile figures being 300 and 42,000 IGPD/ft (50 and 7200 mm^2/s).

   The transmissivity of the bedrock aquifer was estimated using three different approaches. Calculations were carried out using pumping test data, well record data and calculated ground water flow rates. The average transmissivity values found by the three methods were 3,500, 3,200 and 6,950 IGPD/ft (600, 550 and 1200 mm^2/s) respectively. The 10 and 90 percentile figures for the bedrock transmissivity determined from well record data were 610 and 17,000 IGPD/ft (105 and 3000 mm^2/s), a significantly smaller range than was found for the overburden wells.

   No significant differences in the transmissivity values were found when the bedrock wells were divided into classes

according to the static water levels. It was found, however, that the wells where the aquifer was shallow showed a significantly larger range in transmissivity than the deeper wells. This supports the idea of shallow development of karst solution channels and conduits.

5) The extent of the ground water basin was estimated from the piezometric map and was found to have an area of 26.4 square miles (68.5 km^2) which is 13 per cent smaller than the topographic basin area (30.2 square miles or 77.3 km^2). The small difference between the topographic and ground water basin areas supports the theory that the karst is not well developed.

6) The bedrock aquifer was found to be recharged at about 10 inches (254 mm) per year, primarily by the overburden aquifer. Discharge from the bedrock aquifer is at discrete springs along the creeks in a number of swamps, and along portions of the creek beds.

7) The surface runoff component of the streamflow was separated from the ground water discharge component using conventional baseflow separation techniques. The baseflow was found to average 14.0 cfs (0.40 m^3/s) during 1966-69 at the federal gauge, which is approximately 68 per cent of the streamflow of 20.6 cfs (0.58 m^3/s). The average baseflow was equivalent to an average annual ground water runoff rate of 9.8 inches/yr (250 mm/yr) from the area of the basin.

The relatively high degree of consistency of the recession curve determined from the stream hydrographs, indicates that the majority of the ground water discharge comes from one principal aquifer or several aquifers having similar hydraulic characteristics.

8) The evaluation of the specific yield in the different sub-basins during periods when recharge by precipitation and evapotranspiration are assumed negligible, can be used as a measure of the characteristics of the aquifers feeding the streams. The yields vary from 7.2 per cent in the Boy Scouts Camp sub-basin, to 12.7 per cent in the Eden Mills sub-basin, indicating that the yields in the upstream part of the Blue Springs Creek basin are smaller than in the lower sections. The higher yields were related to large areas of thick persistent kame, esker and outwash sands and gravels, which occur principally in the lower sections of the basin.

9) The steady state piezometric surface was successfully modelled in the basin using finite difference techniques. In a total piezometric relief of 280 feet (85 metres), the heads were simulated to an average error between the modelled and measured heads of 6.8 feet (2.1 metres) with the largest error being 28 feet (8.5 metres). The transmissivities used in the model were 1.5 times larger than the median transmissivities found by well analysis.

10) Incorporating variations with time, the ground water flow regime was modelled on a monthly basis from 1966 to 1972. The variations in the heads at the six observation wells in the basin were modelled accurately, with the mean difference between the modelled and measured head variations being 0.95 feet (0.3 metres). The storage coefficients throughout the

basin ranged from 0.04 to 0.16. The recharge to the ground water, determined by water budget considerations, was verified by the model.

11) Experience with the ground water models showed that detailed input data are required to simulate the piezometric heads satisfactorily. It was found that the results of the models are sensitive to changes in transmissivity, coefficient of storage and ground water recharge values used in the model. It was difficult to generalize the parameters in the Blue Springs Creek representative basin, as all of the sensitive parameters were found to vary considerably across the basin; the ground water recharge also varies significantly with time. The only data not sensitive to the model operation are the river permeabilities; high values can generally be used for the modelling. Thus, in examining the the feasibility of extrapolating the Blue Springs Creek model to other areas of similar physiography, basic data on the three sensitive input parameters would be required for the area to be modelled.

12) Thirty-four of a total of 73 ground water and surface water samples were found to have a total dissolved solids concentration above the permissible limit (500 mg/l) recommended for drinking water. All except one of the samples were found to be very hard, although much of the hardness appears temporary and could be reduced easily by suitable home treatment facilities.

13) The majority of the water was found to be of the calcium-magnesium and bicarbonate-chloride-sulphate facies type. Two samples were found to be of different hydrochemical facies; one is likely due to cation exchange through the clays and the other possibly due to contamination by fertilizers or sewage.

The relative degree of metamorphism of the water was found to vary across the basin with the lowest metamorphism occurring near the recharge and discharge zones. Most of the water samples are in the first or second stage of metamorphism.

14) The ground water was generally found to be saturated with respect to calcite and undersaturated with respect to dolomite. Surface waters were often supersaturated with respect to both minerals. The surface water would therefore tend to precipitate calcite and dolomite.

15) The calcium to magnesium ratios (Ca/Mg) of the waters were compared to the (Ca/Mg) of the aquifer materials. It was found that the waters had ratios comparable to the overburden aquifer material, even the samples from the bedrock wells. The water infiltrating down through the overburden becomes saturated with calcite and when the water enters the bedrock it is unable to equilibriate with the dolomite, due to the deposition of calcite in the aquifer material. Thus the water never becomes saturated with respect to dolomite and never approaches the (Ca/Mg) of the dolomite.

The areal trends in the ratios were investigated. Flow patterns of the ground water can be inferred from the (Ca/Mg) trends; the flow pattern was found to match, in a general way, the pattern found by consideration of the piezometric surface.

16) The trends in water chemistry match the ground water flow patterns and ground water divide location better than they match the topographic flow patterns and topographic basin divide, confirming the ground water flow patterns found by consideration of the piezometric surface.

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APPENDIX 1

Water Budget in the Blue Springs Creek Basin, 1966-1972

A summary of this Appendix is given in Table 6.

Year	Month	P	SP	W	E	S	B	O	MWH	ΔGWS	SS	PC	R	LSS
1966	Sept.								23.47	10.00				10.00
	Oct.	1.49	0.0	1.49	1.56	0.34	0.30	0.04	24.14	- 8.04	9.81	0.0	-0.02	10.02
	Nov.	5.59	0.0	5.59	0.68	0.59	0.38	0.21	22.84	15.62	10.0	0.59	1.00	9.61
	Dec.	4.59	0.5	4.09	0.2	1.19	0.51	0.68	20.78	24.7	10.0	3.21	1.49	11.33
1967	Jan.	2.77	0.0	3.27	0.1	0.88	0.71	0.17	20.13	7.8	10.0	3.00	1.02	13.31
	Feb.	1.87	0.0	1.87	0.1	0.93	0.77	0.16	20.43	- 3.6	10.0	1.61	0.62	14.30
	March	1.07	0.0	1.07	0.2	1.24	0.84	0.40	18.76	20.04	10.0	0.47	1.64	13.13
	April	3.87	0.0	3.78	1.0	3.48	1.41	2.07	17.42	16.08	10.0	0.71	2.05	11.79
	May	1.86	0.0	1.86	1.78	1.51	1.05	0.46	18.64	-14.64	9.62	0.0	0.46	11.33
	June	7.22	0.0	7.22	3.91	1.43	0.89	0.54	18.33	3.72	10.0	2.39	1.04	12.68
	July	2.95	0.0	2.95	3.65	1.56	1.08	0.48	19.04	- 8.52	8.82	0.0	0.74	11.94
	Aug.	2.92	0.0	2.92	3.40	0.80	0.63	0.17	20.51	-17.64	8.17	0.0	-0.07	12.01
	Sept.	2.78	0.0	2.78	2.68	0.65	0.52	0.13	21.02	- 6.12	8.14	0.0	0.27	11.74
	TOTAL	38.89	0.5	38.89	19.26	14.60	9.09	5.51		29.40		11.98	10.24	
1967	Oct.	3.78	0.0	3.78	1.60	0.94	0.71	0.23	20.44	6.96	10.0	0.09	0.99	10.84
	Nov.	2.10	0.0	2.10	0.5	1.04	0.78	0.26	20.05	4.68	10.0	1.34	0.96	11.22
	Dec.	3.35	0.0	3.35	0.2	1.38	1.06	0.33	19.05	12.0	10.0	2.82	1.54	12.50
	Jan.	3.54	0.7	2.84	0.1	1.02	0.92	0.10	19.81	- 9.12	10.0	2.64	0.55	14.59
	Feb.	1.65	0.3	2.05	0.1	1.91	0.99	0.92	18.88	11.16	10.0	1.03	1.42	14.20
	March	2.30	0.0	2.60	0.2	3.07	1.31	1.76	17.42	17.52	10.0	0.64	2.01	12.83
	April	1.34	0.0	1.34	1.5	2.03	1.40	0.62	17.98	- 6.72	9.22	0.0	1.13	11.70
	May	3.02	0.0	3.02	1.95	1.34	0.96	0.38	18.90	-11.04	9.91	0.0	0.52	11.18
	June	2.85	0.0	2.85	3.25	0.92	0.67	0.25	19.81	-10.92	9.26	0.0	0.23	10.95
	July	2.52	0.0	2.52	3.85	0.63	0.45	0.18	20.82	-12.12	7.75	0.0	-0.03	10.98
	Aug.	7.35	0.0	7.35	3.80	1.09	0.60	0.49	20.34	5.76	10.0	0.81	0.83	10.96
	Sept.	3.75	0.0	3.75	3.13	1.04	0.73	0.32	20.44	- 1.20	10.0	0.30	0.68	10.58
	TOTAL	37.55	1.0	37.55	20.18	16.41	10.58	5.84		6.96		9.67	10.83	

Year	Month	P	SP	W	E	S	B	O	MWH	ΔGWS	SS	PC	R	LSS
1968	Oct.	2.53	0.0	2.53	1.92	0.82	0.65	0.17	20.75	- 3.72	10.0	0.44	0.50	10.52
	Nov.	3.90	0.0	3.90	0.33	1.22	0.75	0.47	19.82	11.16	10.0	3.10	1.19	12.43
	Dec.	3.45	0.7	2.75	0.2	1.59	1.15	0.44	19.54	3.36	10.0	2.11	1.28	13.26
1969	Jan.	2.57	0.5	2.77	0.1	1.34	1.11	0.23	18.65	10.68	10.0	2.44	1.53	14.17
	Feb.	0.93	0.4	1.03	0.1	1.61	1.25	0.36	19.27	- 7.44	10.0	0.57	0.95	13.79
	March	1.89	0.1	2.19	0.2	2.38	1.69	0.69	17.45	21.84	10.0	1.30	2.56	12.53
	April	3.95	0.0	4.05	1.23	3.42	1.95	1.47	16.63	9.84	10.0	1.35	2.34	11.54
	May	2.93	0.0	2.93	2.30	2.10	1.36	0.74	17.50	-10.44	9.89	0.0	0.94	10.60
	June	1.72	0.0	1.72	3.16	1.25	0.96	0.29	19.12	-19.44	8.16	0.0	0.18	10.42
	July	3.49	0.0	3.49	4.00	0.89	0.69	0.20	20.24	-13.44	7.45	0.0	0.15	10.27
	Aug.	3.28	0.0	3.28	4.10	0.80	0.51	0.29	21.22	-11.76	6.34	0.0	0.04	10.23
	Sept.	0.59	0.0	0.59	3.02	0.44	0.38	0.06	22.32	-13.2	3.85	0.0	-0.14	10.37
	TOTAL	31.23	1.7	31.23	20.66	17.86	12.45	5.41		-22.56		11.31	11.52	
1969	Oct.	2.86	0.0	2.86	1.57	0.47	0.34	0.13	22.19	1.56	5.01	0.0	0.40	9.97
	Nov.	4.05	0.2	3.85	0.46	0.84	0.53	0.31	21.68	6.12	8.09	0.0	0.77	9.20
	Dec.	2.24	0.9	1.54	0.2	0.61	0.48	0.13	22.09	- 4.92	9.30	0.0	0.28	8.92
1970	Jan.	1.23	1.2	0.93	0.1	0.48	0.44	0.04	22.31	- 2.64	10.0	0.09	0.33	8.68
	Feb.	0.85	1.2	0.85	0.1	0.53	0.43	0.10	22.35	- 0.48	10.0	0.65	0.41	8.92
	March	2.10	0.3	3.00	0.2	0.97	0.54	0.43	20.26	25.08	10.0	2.37	1.54	9.75
	April	3.00	0.0	3.30	1.13	2.55	1.34	1.21	18.46	21.60	10.0	0.96	2.20	8.51
	May	2.27	0.0	2.27	2.47	1.39	0.93	0.46	19.29	- 9.96	9.34	0.0	0.53	7.98
	June	2.14	0.0	2.14	3.45	0.78	0.65	0.13	20.59	-15.60	7.90	0.0	0.03	7.95
	July	5.42	0.0	5.42	4.08	0.72	0.52	0.20	20.96	- 4.44	9.04	0.0	0.34	7.61
	Aug.	2.52	0.0	2.52	3.90	0.54	0.43	0.11	21.83	-10.44	7.55	0.0	0.01	7.60
	Sept.	4.04	0.0	4.04	3.00	0.65	0.45	0.20	21.94	- 1.32	8.39	0.0	0.39	7.21
	TOTAL	32.72	3.8	32.72	20.66	10.53	7.08	3.45		4.56		4.07	7.23	

Year	Month	P	SP	W	E	S	B	O	MWH	ΔGWS	SS	PC	R	LSS
1970	Oct.	3.69	0.0	3.69	1.95	0.80	0.54	0.26	21.69	3.0	9.87	0.0	0.66	6.55
	Nov.	2.23	0.0	2.23	0.57	1.19	0.70	0.49	21.11	6.96	10.0	1.04	0.97	6.62
	Dec.	3.85	0.9	2.95	0.2	1.49	0.89	0.60	20.27	10.08	10.0	2.15	1.29	7.48
1971	Jan.	2.06	1.2	1.76	0.1	0.99	0.81	0.18	20.26	0.12	10.0	1.48	0.81	8.15
	Feb.	3.02	1.1	3.12	0.1	1.02	0.78	0.24	20.26	0.0	10.0	2.78	0.78	10.15
	March	1.61	0.8	1.91	0.2	1.62	1.23	0.39	19.37	10.68	10.0	1.32	1.65	9.82
	April	1.32	0.0	2.12	0.67	3.58	1.63	1.95	17.27	25.2	9.5	0.0	2.63	7.19
	May	1.34	0.0	1.34	2.25	1.49	1.14	0.35	18.85	-18.96	8.24	0.0	0.38	6.81
	June	4.85	0.0	4.85	3.62	1.34	0.75	0.59	19.30	- 5.4	8.88	0.0	0.53	6.28
	July	4.61	0.0	4.61	3.70	0.86	0.49	0.37	20.34	-12.48	9.42	0.0	0.0	6.28
	Aug.	6.10	0.0	6.10	3.69	1.03	0.57	0.46	20.24	1.2	10.0	1.37	0.61	7.04
	Sept.	1.90	0.0	1.90	3.35	0.88	0.51	0.37	20.88	- 7.8	8.18	0.0	0.19	6.85
	TOTAL	36.58	4.0	36.58	20.40	16.29	10.04	6.25		12.60		10.14	10.50	
1971	Oct.	1.49	0.0	1.49	2.52	0.51	0.37	0.14	21.36	- 5.64	7.01	0.0	0.15	6.70
	Nov.	2.00	0.2	1.80	0.74	0.54	0.33	0.21	21.87	- 6.12	7.86	0.0	0.09	6.61
	Dec.	3.93	0.6	3.53	0.2	1.16	0.57	0.59	20.87	12.0	10.0	0.60	1.05	6.26
1972	Jan.	2.84	0.5	2.94	0.1	1.04	0.89	0.15	20.52	4.2	10.0	2.69	1.06	7.89
	Feb.	2.75	1.1	2.15	0.1	0.94	0.83	0.11	19.55	11.64	10.0	1.94	1.30	8.53
	March	4.71	0.9	4.91	0.2	1.16	0.92	0.24	18.89	7.92	10.0	4.47	1.24	11.76
	April	2.57	0.0	3.47	0.75	4.85	1.77	3.08	16.27	31.44	9.64	0.0	3.02	8.74
	May	3.68	0.0	3.68	2.70	2.18	1.45	0.73	17.59	-15.84	9.89	0.0	0.72	8.02
	June	4.59	0.0	4.59	3.00	1.35	0.83	0.52	18.47	-10.56	10.0	0.96	0.41	8.57
	July	2.22	0.0	2.22	3.90	1.05	0.68	0.37	19.41	-11.28	7.95	0.0	0.23	8.34
	Aug.	2.00	0.0	2.00	3.71	0.67	0.58	0.09	20.84	-17.16	6.15	0.0	-0.10	8.44
	Sept.	3.12	0.0	3.12	3.03	0.57	0.49	0.08	21.34	- 6.00	6.16	0.0	0.25	8.09
	TOTAL	35.90	3.3	35.90	20.95	16.02	9.71	6.31		- 5.4		10.66	9.42	

P Precipitation in inches
SP Snow pack storage at end of month, in inches of water equivalent
W Water available above soil, in inches
E Calculated evapotranspiration, in inches
S Streamflow, in inches
B Baseflow, estimated from hydrograph, in inches
O Overland flow, in inches
MWH Mean well level in the six observation wells at end of month, in feet below ground surface
ΔGWS Change in ground water storage from end of previous month to the end of the month in inches
SS Soil storage level at the end of the month, in inches; initial level assumed
PC Percolation, in inches
R Ground water recharge, in inches
LSS Relative lower soil storage level, in inches; initial level assumed

APPENDIX 2

Chemical Composition of the Water Samples

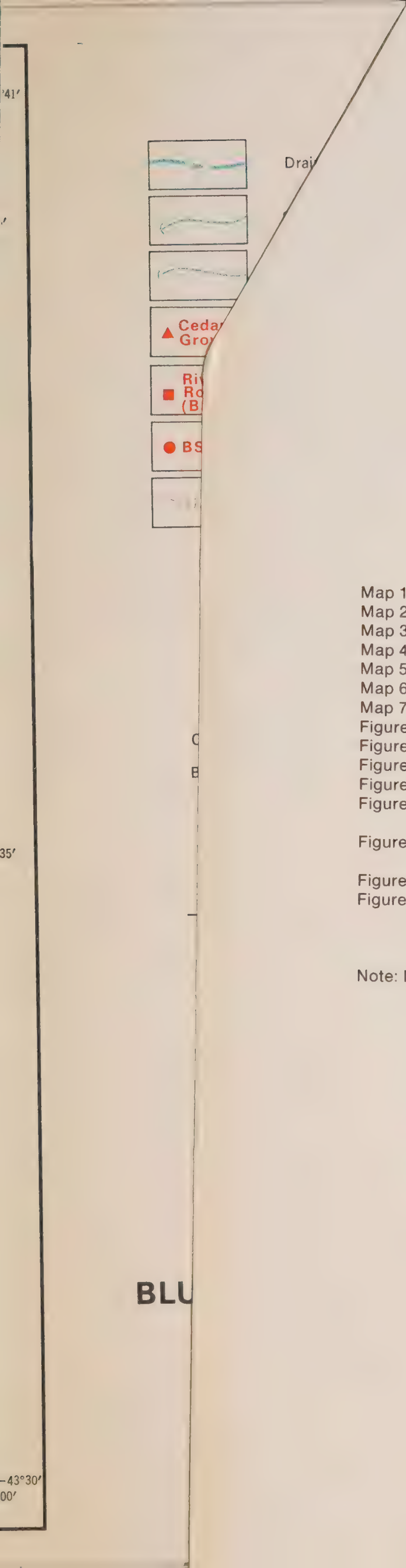
in the Blue Springs Creek Basin

Sample No	Total Concentration in milli-moles/litre								Lab	Degree of Saturation with respect to		Total Dissolved Solids	Iron	Hardness as mg/l
	Ca^{++}	Mg^{++}	HCO_3^-	SO_4^{--}	NA^+	K^+	CL^-	NO_3^-	pH	Calcite	Dolomite	mg/l	mg/l	$CaCO_3$
01	1.92	1.11	4.96	0.42	0.13	0.04	0.20	0.01	7.60	0.04	- 0.11	458.5	.1	303
02	1.97	1.11	5.20	0.36	0.30	0.10	0.40	0.03	7.40	- 0.12	- 0.45	484.7	.05	308
03	2.87	1.40	6.36	0.73	0.96	0.33	0.68	0.23	7.30	- 0.02	- 0.30	679.7	.05	427
04	2.37	1.03	5.04	0.80	0.17	0.20	0.25	0.05	7.40	- 0.08	- 0.47	528.5	.1	340
05	4.05	1.19	6.64	0.97	2.00	0.04	1.72	0.40	7.20	0.02	- 0.44	822.4	.05	524
06	3.80	0.82	5.82	0.80	1.04	0.03	1.02	0.31	7.20	- 0.04	- 0.70	684.1	.1	462
07	2.30	0.82	4.88	0.41	0.17	0.01	0.42	0.12	7.50	0.01	- 0.38	475.2	.05	312
08	1.30	0.86	4.24	0.16	0.30	0.03	0.06	0.00	7.70	- 0.06	- 0.25	356.9	.65	216
09	1.97	0.74	4.60	0.27	0.22	0.03	0.48	0.01	7.90	0.33	0.28	427.3	.15	271
10	2.32	1.03	5.24	0.68	0.17	0.02	0.23	0.00	7.40	- 0.07	- 0.44	515.5	.05	335
11	2.05	1.07	5.52	0.26	0.09	0.03	0.14	0.02	7.50	0.02	- 0.20	478.7	.1	311
12	2.35	1.11	5.20	0.42	0.30	0.14	0.37	0.16	7.30	- 0.16	- 0.60	513.4	.05	346
13	2.62	0.91	4.88	0.35	0.13	0.09	0.28	0.23	7.30	- 0.13	- 0.68	489.2	.05	352
14	2.50	0.95	4.98	0.40	0.39	0.04	0.82	0.11	7.60	0.15	- 0.08	511.0	.45	344
15	2.17	0.95	4.88	0.49	0.39	0.04	0.40	0.06	7.50	- 0.01	- 0.35	482.8	.05	312
16	3.32	1.19	6.08	0.69	0.91	0.10	0.87	0.27	7.20	- 0.08	- 0.55	671.6	.05	451
17	2.20	0.99	5.04	0.44	0.35	0.05	0.48	0.05	7.40	- 0.10	- 0.49	491.2	.45	318
18	1.70	0.74	4.12	0.29	0.30	0.03	0.45	0.00	7.70	0.03	- 0.26	389.8	.30	244
19	2.30	1.07	5.32	0.42	0.35	0.08	0.62	0.05	7.60	0.14	- 0.01	518.8	.05	336
20	2.32	1.11	5.20	0.59	0.30	0.05	0.54	0.05	7.40	- 0.07	- 0.41	524.6	.02	343
21	1.80	0.99	5.00	0.28	0.13	0.05	0.25	0.00	7.60	0.03	- 0.16	441.6	1.5	278
22	1.87	1.11	4.68	0.30	0.22	0.03	0.28	0.00	7.60	0.01	- 0.15	432.5	.1	298
23	1.98	0.82	5.16	0.02	0.17	0.03	0.28	0.00	7.60	0.09	- 0.16	430.7	.85	279
24	3.27	1.40	7.52	0.36	0.30	0.11	0.42	0.00	7.10	- 0.09	- 0.49	684.9	.05	467
25	2.12	0.95	4.88	0.42	0.26	0.06	0.37	0.06	7.60	0.08	- 0.15	470.8	.60	310
26	2.12	0.86	4.72	0.35	0.13	0.03	0.20	0.09	7.50	- 0.03	- 0.40	444.6	.05	298
27	2.52	0.82	4.96	0.64	0.61	0.18	0.51	0.23	7.50	0.04	- 0.35	537.4	.15	340
28	2.20	0.86	5.10	0.37	0.17	0.03	0.31	0.05	7.50	0.01	- 0.33	475.1	.05	308
29	1.85	0.82	4.62	0.30	0.30	0.03	0.42	0.01	7.90	0.30	0.30	428.6	.10	270
30	2.02	1.15	4.48	0.40	0.83	0.28	0.71	0.32	7.50	- 0.08	- 0.36	495.1	.02	318
31	2.92	1.23	5.16	1.62	0.26	0.22	0.23	0.02	7.50	0.08	- 0.16	641.1	.02	416
32	2.85	1.85	6.96	0.99	0.74	2.05	0.79	0.55	7.30	- 0.01	- 0.15	837.1	.05	468
33	2.37	1.07	5.32	0.47	0.17	0.03	0.28	0.07	7.70	0.25	0.20	509.6	.02	346
34	2.47	1.23	5.84	0.51	0.57	0.77	1.30	0.13	7.30	- 0.10	- 0.46	631.3	.02	372
35	1.75	1.03	4.76	0.40	0.13	0.03	0.11	0.00	7.60	- 0.01	- 0.20	431.2	1.1	278

Sample No	Total Concentration in milli-moles/litre Ca^{++}	Mg^{++}	HCO_3^-	SO_4^{--}	NA^+	K^+	CL^-	NO_3^-	Lab pH	Degree of Saturation with respect to Calcite	Dolomite	Total Dissolved Solids mg/l	Iron mg/l	Hardness as mg/l $CaCO_3$
36	2.45	1.28	6.00	0.51	0.48	0.04	0.51	0.16	7.70	0.30	0.37	583.9	.02	372
37	1.45	0.95	4.34	0.30	0.26	0.03	0.11	0.00	7.60	- 0.11	- 0.37	385.9	.85	240
38	1.42	0.86	4.68	0.02	0.26	0.02	0.03	0.00	7.60	- 0.08	- 0.33	373.2	1.3	228
39	2.37	0.74	5.20	0.28	0.13	0.01	0.54	0.02	7.50	0.06	- 0.35	480.7	.05	314
40	2.60	1.03	5.76	0.34	1.00	0.03	1.72	0.02	7.40	0.02	- 0.32	599.6	.05	364
41	3.12	1.52	5.72	0.62	0.39	0.41	0.59	0.55	7.40	0.07	- 0.12	650.7	.05	464
42	3.72	2.26	9.43	3.00	1.22	6.38	2.46	0.73	7.40	0.25	0.35	1474.7	.1	600
43	1.92	1.03	5.08	0.49	0.30	0.24	0.34	0.01	7.50	- 0.05	- 0.32	487.2	.2	296
44	2.75	1.11	4.98	0.57	0.70	0.08	0.79	0.36	7.60	0.17	- 0.00	564.9	.1	386
45	3.20	1.93	8.50	0.49	0.65	0.74	0.99	0.24	7.10	- 0.07	- 0.30	834.1	.05	512
46	2.47	1.11	5.46	0.58	0.61	0.03	1.04	0.01	7.30	- 0.13	- 0.56	567.9	.1	360
47	1.80	0.90	4.64	0.30	0.26	0.03	0.42	0.01	8.20	0.58	0.90	428.7	.1	270
48	1.92	1.07	4.76	0.35	0.74	0.03	1.07	0.01	7.70	0.12	0.03	484.3	.05	300
49	1.80	1.07	4.78	0.29	0.17	0.05	0.23	0.05	7.90	0.30	0.42	434.1	.05	288
50	1.77	1.36	5.44	0.51	0.17	0.31	0.23	0.00	7.50	- 0.06	- 0.19	508.9	.05	314
51	2.15	0.99	4.88	0.33	0.44	0.11	0.59	0.14	7.60	0.08	- 0.13	483.4	.1	314
52	2.35	1.19	5.32	0.69	0.22	0.10	0.40	0.05	7.40	- 0.06	- 0.37	539.6	.05	354
53	1.92	0.95	4.96	0.34	0.17	0.02	0.23	0.00	7.70	0.15	0.03	448.5	.05	288
54	1.95	0.86	4.72	0.29	0.13	0.05	0.14	0.05	7.70	0.14	- 0.03	427.9	.05	280
55	2.25	1.11	5.20	0.30	0.26	0.02	0.90	0.01	7.60	0.12	- 0.01	502.2	.05	334
56	2.57	1.19	6.00	0.50	0.70	0.03	0.40	0.18	7.50	0.13	- 0.03	587.8	1.5	378
57	1.47	0.86	4.32	0.07	0.30	0.02	0.03	0.0	7.70	0.00	- 0.19	359.3	.9	234
58	1.47	0.86	4.36	0.03	0.13	2.02	0.03	0.0	7.60	- 0.09	- 0.37	353.7	.75	234
59	1.82	0.95	4.48	0.04	0.17	0.02	0.54	0.0	7.60	- 0.00	- 0.24	397.0	.6	276
60	0.02	0.02	5.52	1.37	5.75	0.01	0.06	0.0	7.70	- 1.71	- 3.47	604.3	.05	3
61	2.35	1.32	5.16	0.07	0.30	0.20	0.37	0.09	7.50	0.04	- 0.12	481.1	.05	366
62	2.27	1.03	5.28	0.08	0.35	0.03	0.48	0.12	7.40	- 0.05	- 0.40	479.3	.02	332
63	1.92	1.11	4.86	0.12	0.52	0.03	0.68	0.01	8.20	0.63	1.05	450.1	.1	284
64	2.10	0.95	4.80	0.03	0.13	0.04	0.25	0.15	7.50	- 0.02	- 0.33	425.1	.05	304
65	2.15	0.95	4.78	0.37	0.13	0.04	0.25	0.10	7.80	0.27	0.23	456.5	.02	310
66	2.45	1.07	5.36	0.66	0.57	0.56	0.54	0.23	7.40	- 0.04	- 0.40	581.7	.05	352
67	2.97	0.95	5.72	0.48	1.61	0.08	2.43	0.05	7.40	0.06	- 0.33	665.8	.05	392
68	2.22	1.28	5.96	0.34	0.17	0.04	0.23	0.01	7.30	- 0.12	- 0.44	530.2	.02	350
69	2.05	1.11	5.00	0.64	0.22	0.05	0.23	0.02	7.50	- 0.03	- 0.29	491.2	.05	318

Sample No	Total Concentration in milli-moles/litre								Lab	Degree of Saturation with respect to		Total Dissolved Solids	Iron	Hardness
	Ca^{++}	Mg^{++}	HCO_3^-	SO_4^{--}	NA^+	K^+	CL^-	NO_3^-	pH	Calcite	Dolomite	mg/l	mg/l	as mg/l $CaCO_3$
70	2.10	1.19	5.04	0.76	0.44	0.03	0.28	0.05	7.40	- 0.13	- 0.46	517.4	.05	328
71	3.05	1.11	6.32	0.46	0.35	0.06	0.56	0.19	7.20	- 0.08	- 0.55	620.6	.05	416
72	2.85	0.99	6.00	0.66	0.35	0.31	0.65	0.11	7.40	0.06	- 0.28	616.5	.1	396
73	1.85	0.78	4.50	0.27	0.52	0.04	0.40	0.01	8.2	0.58	0.83	421.2	.1	264
81	2.00	0.78	3.81	0.64	0.23	0.04	0.34	0.08	8.0	0.27	0.17	405.6	.05	278
82	1.75	0.98	4.02	0.49	0.18	0.04	0.34	0.05	8.1	0.35	0.49	394.8	.05	272
83	1.82	1.15	4.75	0.35	0.61	0.03	0.86	0.11	7.9	0.26	0.40	430.9	.05	298

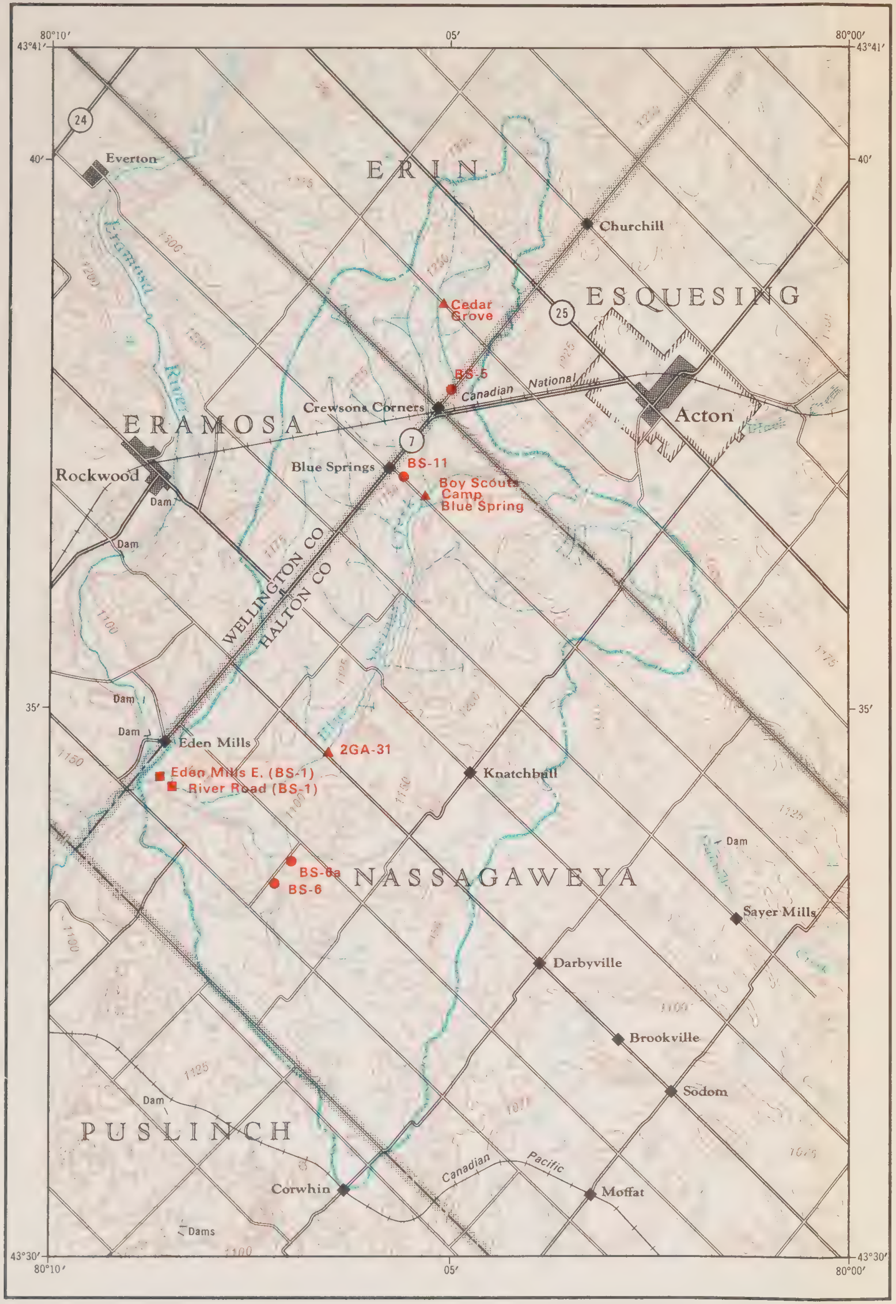
The degree of saturation with respect to calcite and dolomite is described in section 5.4. The location and source of the water samples are shown in Figure 43.



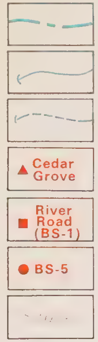
WATER RESOURCES REPORT 10

Maps and Figures in Pocket:

Note: Figures 14/15 and figures 44/45/46 are printed together on two sheets



LEGEND



Drainage basin boundary

Spring forming perennial stream

Spring forming intermittent stream

Recording hydrometric station with station name or number

Hydrometric station, daily readings with station name and number

Hydrometric station, monthly readings with station number

Topographic contour, interval 25 feet

SOURCES OF INFORMATION

Cartography by R. Zimmermann and C. Lochan

Base map derived from 1:50,000 sheets of the National Topographic series.

To accompany Water Resources Report 10

Ontario

MINISTRY OF THE ENVIRONMENT

Water Resources Branch

INTERNATIONAL HYDROLOGICAL DECADE

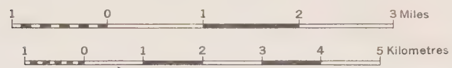
BLUE SPRINGS CREEK DRAINAGE BASIN

MAP 1

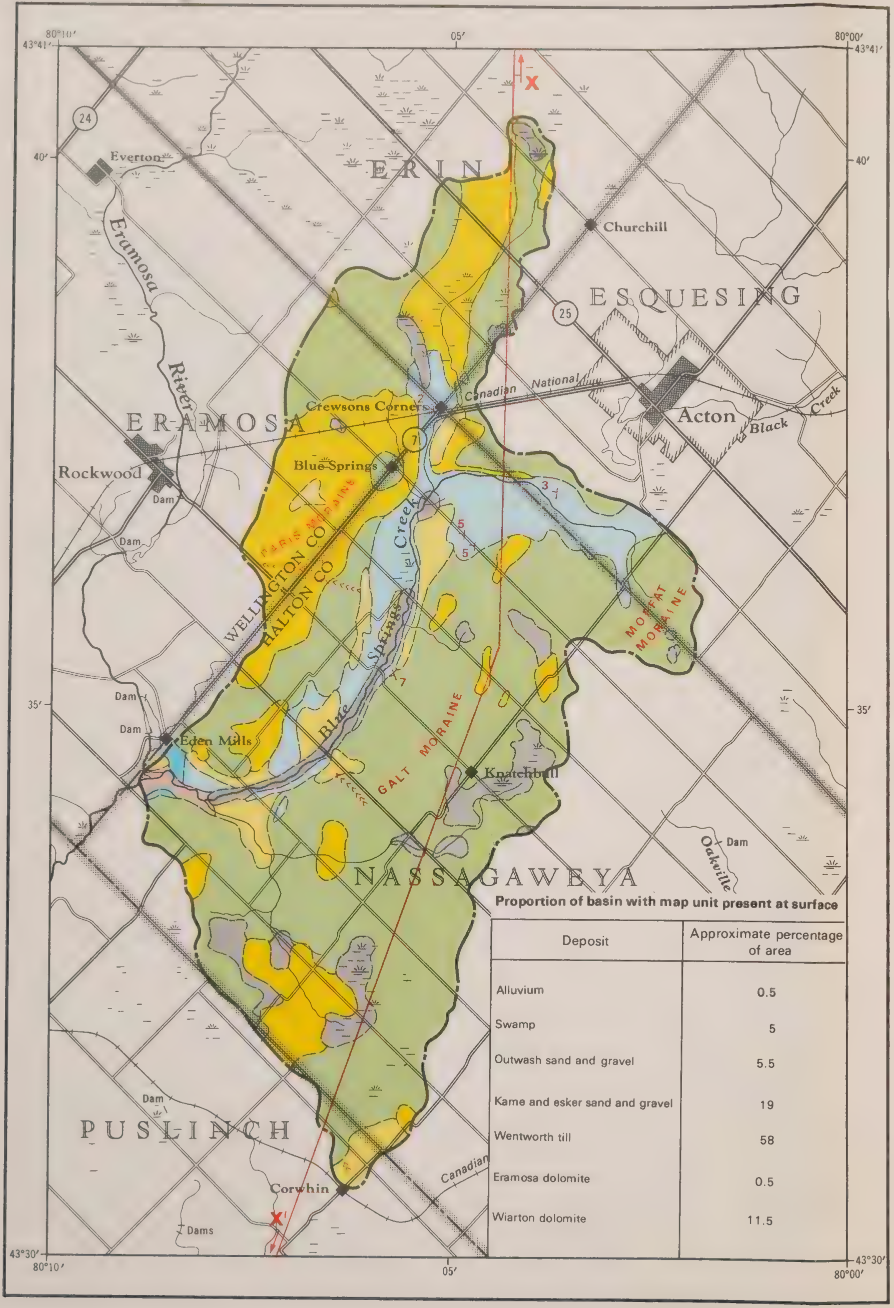
TOPOGRAPHY, DRAINAGE AND HYDROMETRIC STATIONS

Scale 1:100,000

1 inch equals 1.58 miles



MAP 2161



Proportion of basin with map unit present at surface

Deposit	Approximate percentage of area
Alluvium	0.5
Swamp	5
Outwash sand and gravel	5.5
Kame and esker sand and gravel	19
Wentworth till	58
Eramosa dolomite	0.5
Wiarton dolomite	11.5

LEGEND

CENOZOIC
PLEISTOCENE
RECENT

- **Alluvial deposits:** stratified silt, sand and gravel
- **Swamp deposits:** peat, muck, marl

WISCONSINAN

- **Outwash deposits:** stratified sands and gravels
- **Kames and eskers:** sands and gravels
- **Wentworth till:** buff or pink boulder clay, clay or sandy clay

PALEOZOIC
MIDDLE SILURIAN
AMABEL FORMATION

- **Eramosa member:** bituminous dolomite
- **Wiarton member:** bright grey mottled dolomite

Note: The Colpoy Bay member probably outcrops along the middle portion of Blue Springs Creek, but was not identified in the basin.

SYMBOLS

- Geological boundary, approximate
- Esker
- Measured bedrock strike and dip in degrees
- Line of section shown in figure 3

SOURCES OF INFORMATION

Surficial geology compiled by J. Coward from the following references:

Karrow, P.F., 1968, Pleistocene geology of the Guelph area; Ontario Department of Mines, Map 2153.

Sanford, B.V., 1969, Geology of the Toronto-Windsor area; Geological Survey of Canada, Map 1263A.

Cartography by R. Zimmermann and C. Lochan.

Base map derived from 1:50,000 sheets of the National Topographic series.

To accompany Water Resources Report 10

Ontario

MINISTRY OF THE ENVIRONMENT

Water Resources Branch

INTERNATIONAL HYDROLOGICAL DECADE

BLUE SPRINGS CREEK DRAINAGE BASIN

MAP 2

SURFICIAL GEOLOGY

Scale 1:100,000

1 inch equals 1.58 miles

1 0 1 2 3 Miles

1 0 1 2 3 4 5 Kilometres

Map 2141

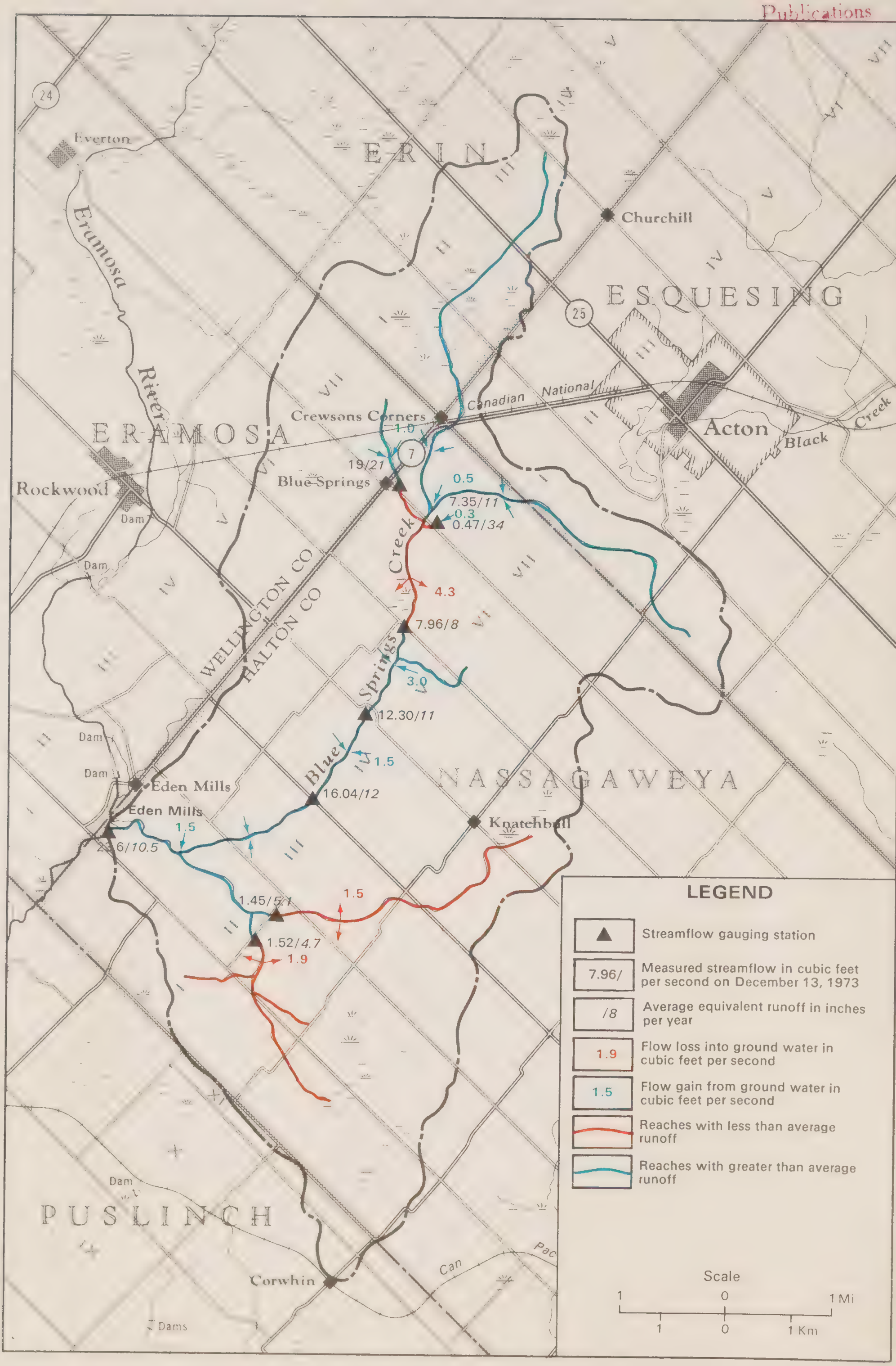


Figure 14. Streamflows in the Blue Springs Creek basin used to determine the losing and gaining reaches along the creek.

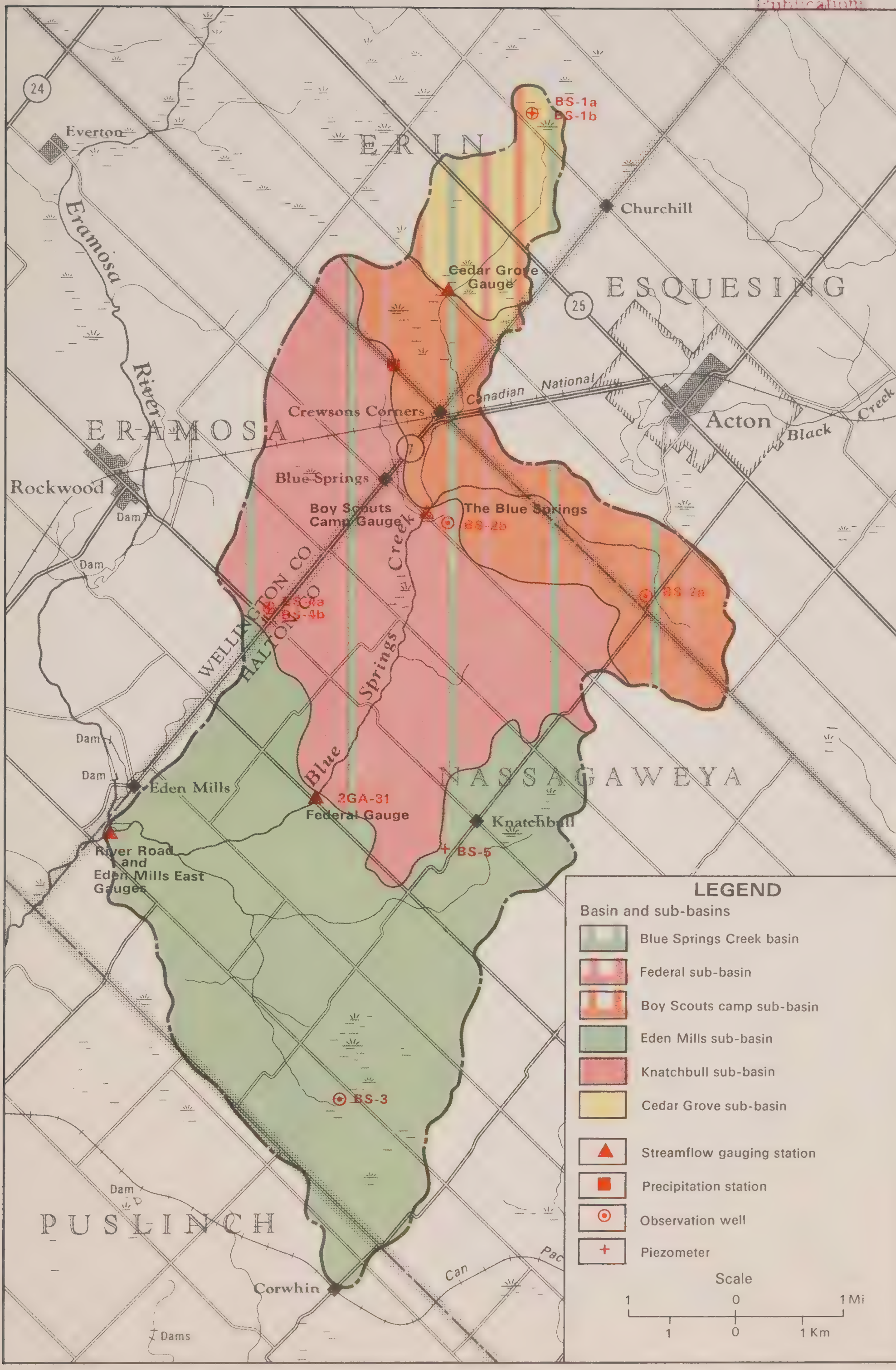
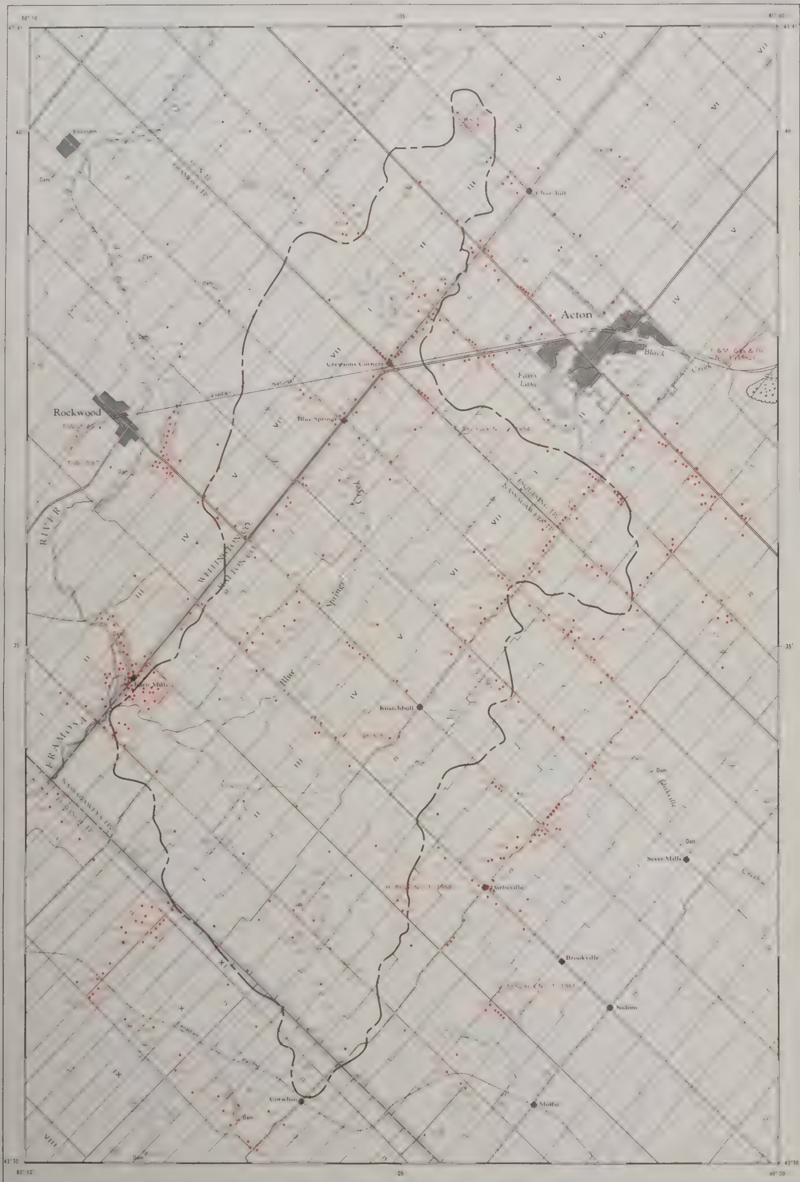


Figure 15. Locations of streamflow stations, observation wells, precipitation stations, piezometers and sub-basins in the Blue Springs Creek basin.



LEGEND

- Water well in overburden
- Water well in bedrock
- Observation well or piezometer in overburden
- Observation well or piezometer in bedrock
- Test hole for oil or gas
- Water well for which a pumping test has been carried out
- Water well number
- Ministry of the Environment observation well number
- Ministry of the Environment piezometer well number and number of piezometers
- Oil or gas hole name and date

SOURCES OF INFORMATION

This map shows the location of water wells for which records have been filed with the Ontario Ministry of the Environment, and the locations of oil and gas wells for which records have been filed with the Ontario Ministry of Natural Resources.

Cartography by M. Lakin and C. ...han

Base map derived from 1:50,000 maps of the National Topographic series.

To accompany *Water Resources Report 10*

Ontario

MINISTRY OF THE ENVIRONMENT
Water Resources Branch

INTERNATIONAL HYDROLOGICAL DECADE

BLUE SPRINGS CREEK DRAINAGE BASIN

MAP 3

LOCATIONS OF WELLS

Scale 1:50,000
1 inch equals 0.79 miles

Miles

Kilometres

MAP 2164

LEGEND

Bedrock surface contour, interval 25 feet

Water well

Elevation of top of bedrock for selected wells in feet

Bedrock exposure

1125 Topographic contour, interval 25 feet

SOURCES OF INFORMATION

Bedrock topography by J. Coward, 1976, on the basis of water well records on file with the Ontario Ministry of the Environment

References
Karrow, P.F., 1968, Pleistocene geology of the Guelph area, Ontario Department of Mines, Map 2153

Sanford, B.V., 1969 Geology of the Toronto-Windsor area, Geological Survey of Canada, Map 1263A

Cartography by M. Lakin and C. Lochan.

Base map derived from 1:50,000 maps of the National Topographic series

To accompany Water Resources Report 10

Ontario
MINISTRY OF THE ENVIRONMENT
Water Resources Branch

INTERNATIONAL HYDROLOGICAL DECADE

BLUE SPRINGS CREEK DRAINAGE BASIN

MAP 4

BEDROCK TOPOGRAPHY

Scale 1:50,000
1 inch equals 0.79 miles

0 ½ 1
Miles

0 ½ 1 2
Kilometres

LEGEND

[illegible] Line of equal [illegible] interval 25 feet

Water well

Overburden [illegible] feet

Bedrock exposure

Topographic contour, interval 25 feet

SOURCES OF INFORMATION

Overburden [illegible] C[illegible]ward 19[illegible] records on file with the Ontario Ministry of the Environment.

References

[illegible] P.F. 1968 P[illegible] geology of the [illegible] Department of Mines Map 2153

Sanford B.V. 1969 Geology of the Toronto-Windsor [illegible] Geological Survey of Canada, Map 1263A

C[illegible] by M[illegible] and C. Lochar[illegible]

Base map derived from 1:50,000 sheets of the National Topog[illegible] series

To accompany Water Resources Report [illegible]

Ontario

MINISTRY OF THE ENVIRONMENT
Water Resources Branch

INTERNATIONAL HYDROLOGICAL DECADE

BLUE SPRINGS CREEK DRAINAGE BASIN

MAP 5

OVERBURDEN THICKNESS

Scale 1:50,000
1 inch equals 0.79 miles

0 ½ 1
Miles

0 ½
Kilometres

MAP 2154

LEGEND

- Soil boundary, approximate
- **Burford and Fox loams** developed from outwash sands and gravels, well drained
- C — **Colwood loam:** developed from silty sand of lacustrine origin, poorly drained
- **Dumfries loam:** developed from coarse textured till material usually on moraines, well drained and often containing many stones and boulders
- **Donnybrook loam** developed from kame and esker sands and gravels, well to excessively drained
- **Farmington loam** developed from very thin till over bedrock
- G — **Guelph loam** developed from tills in till plains or gently sloping moraines, well drained and a good crop soil
- **Killean loam:** imperfectly drained members of the Dumfries loam
- **Lily loam:** poorly drained members of the Dumfries loam
- **Marsh, mesisol and peat.** poorly drained swamp deposits, developed over a variety of parent materials, often indicates ground water discharge areas
- **Parkhill and London loams:** imperfectly and poorly drained members of the Guelph loam

SOURCES OF INFORMATION

Soil configuration compiled by J Coward from the following references

Hoffman, D W, Matthews, B C and Wicklund, R E, 1963, Soil survey of Wellington County, Canada Department of Agriculture and the Ontario Department of Agriculture Report 35

Gillespie, J E Wicklund, R E and Miller M H, 1971, Soils of Halton County, Ontario Department of Agriculture and Food and the Canada Department of Agriculture, Report 43

Cartography by M Lakin and C Lochan

Base map derived from 1 50 000 sheets of the National Topographic series

To accompany Water Resources Report 10

Ontario

MINISTRY OF THE ENVIRONMENT
Water Resources Branch

INTERNATIONAL HYDROLOGICAL DECADE

BLUE SPRINGS CREEK DRAINAGE BASIN

MAP 6

SOILS

Scale 1:50,000
1 inch equals 0.79 miles

0 ½ 1 2
Miles

0 ½ 1 2
Kilometres

LEGEND

Piezometric surface [illegible] interval 25 feet

Elevation of static level for selected wells in feet

Surface water spot elevation in feet

Ground water basin divide, approximate

Ground water discharge area, approximate

Topographic contour interval 25 feet

SOURCES OF INFORMATION

Piezometric surface by J. Coward, 1976, on the basis of water well records on file with the Ontario Ministry of the Environment

Cartography by M. Lakin and C. Lochan

Base map derived from 1:50,000 sheets of the National Topographic series

To accompany Water Resources Report 10

Ontario

MINISTRY OF THE ENVIRONMENT
Water Resources Branch

INTERNATIONAL HYDROLOGICAL DECADE

BLUE SPRINGS CREEK DRAINAGE BASIN

MAP 7

PIEZOMETRIC SURFACE FOR BEDROCK

Scale 1:50,000
1 inch equals 0.79 miles

Miles

Kilometres

MAP 2155

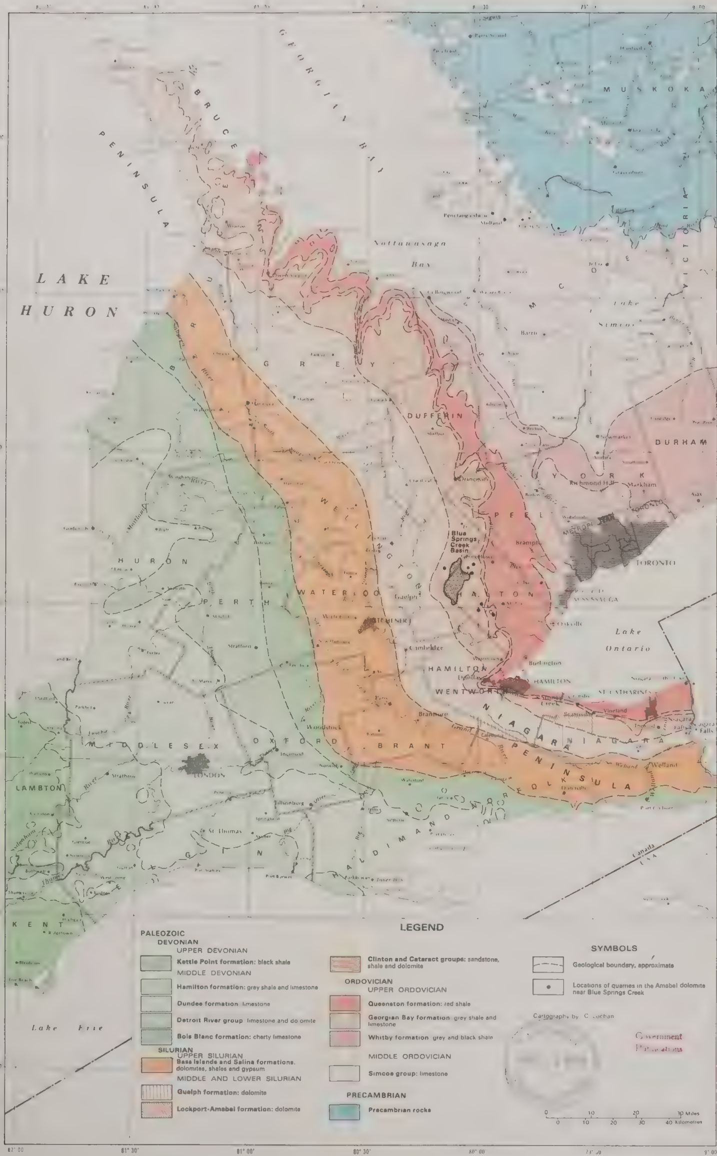


Figure 2. Generalized bedrock geology, southwestern Ontario, after Hewitt (1964, 1972).

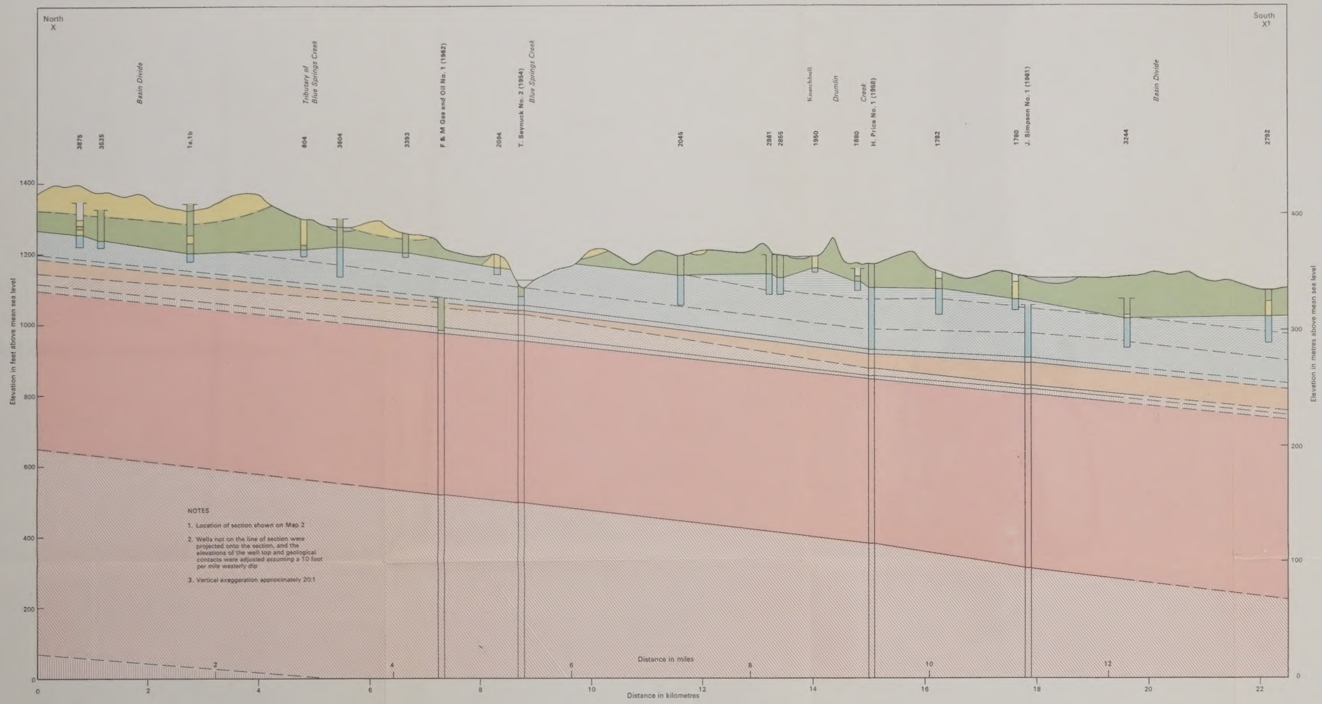


Figure 3. Geologic cross section of bedrock and overburden formations in the Blue Springs Creek basin.

LEGEND

CENOZOIC
PLEISTOCENE
RECENT

Swamp deposits: peat, muck, marl

WISCONSINAN

Till: mainly boulder clay, sand and gravel and clay

Kame: mainly sand and gravel

PALEOZOIC
MIDDLE SILURIAN

Amabel Group: dolomites

The Amabel group was tentatively split into three members on the basis of well log information on the dolomite colour, and on the assumed thickness of each member:

Eramosa member: brown bituminous dolomite

Wiarton member: light grey, grey mottled dolomite

Colpoy Bay member: white to blue-grey dolomite

Reynales formation: brown dolomites and shales

LOWER SILURIAN
CATARACT GROUP

Cabot Head formation: greenish grey and red silty shale

Manitoulin formation: brown and grey dolomite

Whirlpool formation: white and grey sandstone and quartzite

UPPER ORDOVICIAN

Queenston formation: red shale and mudstone with some interbeds of dolomite

Georgian Bay formation: grey and greenish grey shale

Whitby formation: bituminous shale

SYMBOLS

Geological boundary, approximate

Geological boundary, tentative

2045 — Water well number, location on Map 3

H. Price No. 1 (1958) — Oil and gas well number, location on Map 3

SOURCES OF INFORMATION

Geology by J. Coward, 1976, on the basis of water well records on file with the Ontario Ministry of the Environment, and oil and gas well logs assembled by the Ontario Ministry of Natural Resources.

References:

Karrow, P.F., 1968, Pleistocene geology of the Guelph area; Ontario Department of Mines, Map 2153.

Sanford, B.V., 1969, Geology of the Toronto-Windsor area; Geological Survey of Canada, Map 1263A.

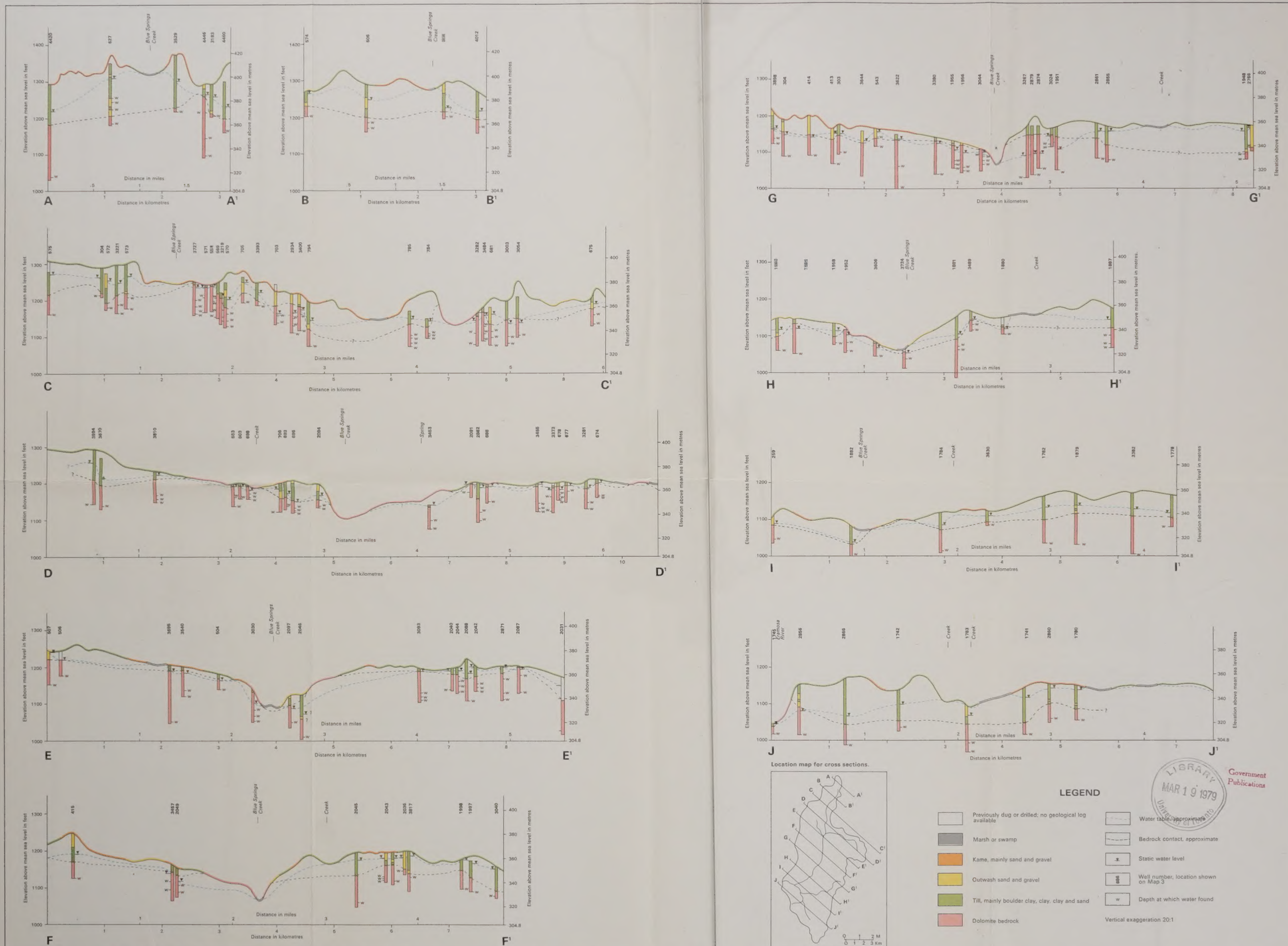


Figure 5. Geologic and hydrogeologic cross sections in the Blue Springs Creek basin.

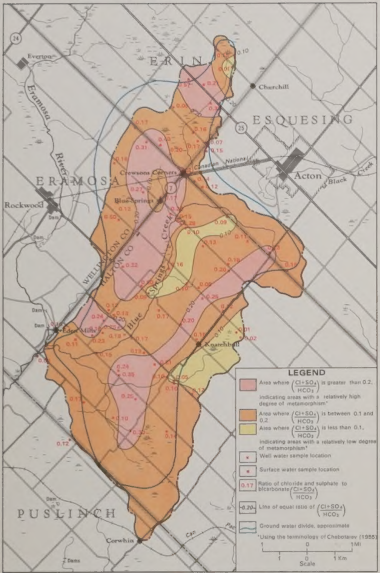


Figure 44. Chloride and sulphate to bicarbonate milli-equivalents per litre ratios in water samples.

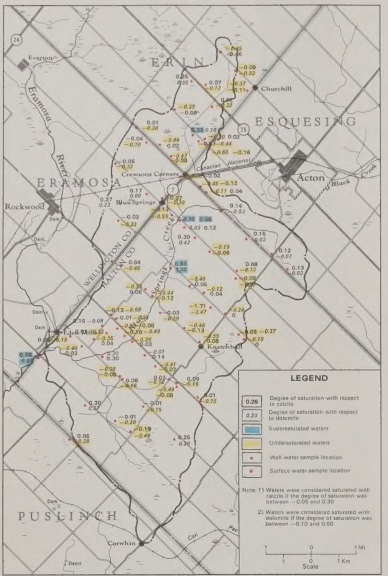


Figure 45. Degree of saturation with respect to dolomite and calcite.

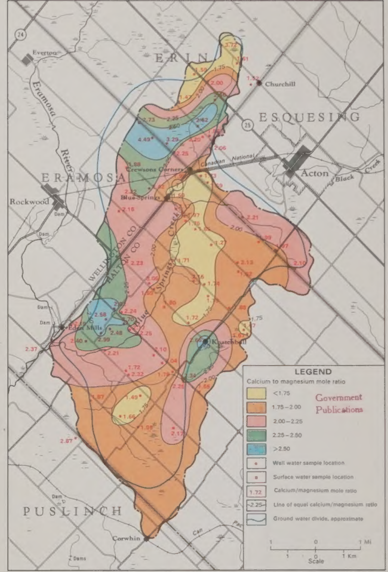


Figure 46. Calcium to magnesium ratios of water samples.